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ERRATA

BRADLEY VOLUME, VOLUME 258-A

p. vii, line 11. Change Frank M. Schairer to J. F. Schairer.

p. 39, line 2. Change Kiever to McIver.

p. 49, 51. Explanations for figs. 2 and 3 are interchanged.

p. 52, 4th line up. Change etxensive to extensive.

p. 87, line 15. Change qaurtzite to quartzite.

p. 107, line 22. Change facies of the anlysis to facets of the analysis.

p. 109, line 21. Change vesicular to vesicular.

p. 117, line 11. Change overriden to overridden.

p. 123, line 21. Change casual to causal.

p. 137, line 14. Change no older to younger.

p. 197, 13th line up. Change strike-dip to strike-slip.

p. 199, line 24. Change Pennsylvania to Pennsylvanian.

p. 203, Insert Pierce, W. G., 1957, Heart Mountain and South Fork detachment thrusts of Wyoming: Am. Assoc. Petroleum Geologists Bull., v. 41, p. 591-626.

p. 245, line 6 (13.1.4). Change Na_3 to Na_2.

p. 288, 5th line up. Change Icacinadceae to Icacinaceae.

p. 289, 5th line up. Change cosatal vegetation bodering to coastal vegetation bordering.

p. 329, 4th line up. Change fosil to fossil.

p. 329, 3d line up. Insert period after Lake.

p. 331, line 13. Change considred to considered.

p. 331, 15th line up. Change moluusks to mollusks.

p. 331, 9th line up. Change *P. spilonotus* to *P. abyssicola*.

p. 333, line 2. Change *Entosphemus* to *Entosphenus*.

p. 333, line 15. Change modern to moderate.

p. 344, line 6 (of abstract). Change 6 to 4.

p. 347, lines 7 and 12. Change 6 to 4.

p. 351, lines 17 and 20. Change intercanyon to intracanyon.

THE BRADLEY VOLUME

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in appreciation

of

WILMOT HYDE BRADLEY

on his sixty-first birthday

BRADLEY VOLUME

1960

NEW HAVEN, CONNECTICUT

I

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FOREWORD

Late in 1958 several of Bill Bradley's friends decided that a Festschrift in his honor was the most suitable means of expressing a widely shared appreciation of what he has done both for geology and for geologists. This plan met an enthusiastic reception, culminating in the present volume. The book has no central theme other than the breadth of geology, as expressed in these brief papers.

There could be no better medium for the publication of a Bradley volume than this Journal, which is so closely associated with Bradley's alma mater and of which he has been an Associate Editor for 15 years.

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W T Bradley

WILMOT HYDE BRADLEY

Geologist, Geomorphologist, Paleolimnologist, Paleontologist, Administrator

Wilmot Hyde Bradley was born April 4, 1899, in Westville, Connecticut, a village near New Haven. As a perceptive and receptive boy growing up in a rural environment, he soon found a life-long interest: natural history.

When he entered Sheffield Scientific School at Yale University, World War I was on and for two years, 1918-19, he was in the "Yale Navy", that is, in the U. S. Naval Reserve Force detachment based at New Haven. At first he was uncertain what his major collegiate subject would be, but settled on chemistry. Geology did not beckon to him until his senior year in 1920. That summer, after graduating with a Bachelor of Philosophy degree, he had his first U. S. Geological Survey assignment: to serve as Field Assistant to F. C. Calkins on a project in the Cottonwood Canyon district of the Wasatch Mountains.

At the end of the field season Bradley returned to Yale to start graduate work and was in residence there during that academic year and the following one. In 1921 he qualified for an appointment on the U. S. Geological Survey as Geologic Aid. At that time the Fuels Branch (then Section of Geology of Fuels), the Chief of which was K. C. Heald, was the largest unit in the Geologic Division (then Branch). During the field seasons of 1921 and 1922, Bradley was assigned to a Fuels Branch party mapping in northwestern Colorado under the leadership of J. D. Sears. Those two field seasons were a stimulating experience. They offered to Bradley his first acquaintance with the mainly lacustrine Green River formation, of middle Eocene age, in which he immediately became interested. Moreover, the alert young geologists of both seasons were learning field techniques and engaged in lively discussions of every aspect of the geology they were seeing. In 1922 Bradley was promoted to Assistant Geologist at a salary of $1,800. After the field season was over he married Catrina van Benschoten, whom he had known for many years, and they moved to Washington where he took up work on oil shale. In 1923, at the age of 24, he had a field party of his own in southwestern Wyoming. His assistant was a Yale classmate, C. H. Dane, with whom he has been closely associated ever since.

Shortly after Bradley came to Washington, David White (then Chief Geologist), who was interested in the composition and origin of coal and oil shale, urged him to study not only the field relations of the oil shale of the Green River formation but also its composition and origin. White turned over to him a set of thin sections and photomicrographs prepared under the direction of C. A. Davis, who had died before he was able to do anything except some preliminary work. This was the beginning of an intensive field and laboratory study of the Green River and its microscopic fossils. Though it has been interrupted, it is still continuing. In 1927 a thesis entitled "Origin and microfossils of the oil shale of the Green River formation of Colorado and

Utah" was submitted in partial fulfillment of the requirements for the degree of Doctor of Philosphy at Yale. It was published in essentially the same form in 1931.

During several depression years in the 1930's, Bradley worked on the Oriskany sandstone of southern New York and its gas possibilities—an undertaking that started as a Public Works Administration project. Later in the same decade he organized and participated in the work of a group of geologists, paleontologists, and others in a study of the Piggot North Atlantic deep-sea cores. The longest interruption to his work in the west began in 1942, when he was designated Geologist in Charge of the Military Geology Unit (later Branch). This assignment lasted until 1944, when he was appointed Chief Geologist. He was the youngest geologist to hold that position and served the longest term. Before he left for Europe in the spring of 1959 to attend the XIV International Limnological Congress in Vienna, he arranged with the Director to be relieved as Chief Geologist, but that was not generally known until he returned in the fall to take up his new duties as Research Geologist.

His publications on the Green River formation are a good example of his meticulous gathering of field data, his imaginative interpretation of geological history, his mastery of pertinent disciplines and techniques, his rigorous logic, and his lucid writing. The Green River is in many ways a remarkable formation, containing potential oil resources far in excess of the combined production up to the present time and the conventional reserves in the United States. Bradley's work led to new concepts of the history of the ancient lakes in which the oil shale and other deposits were laid down and their relations to contemporaneous strata. Under his treatment this basically economic study branched into unexpected avenues, all of which were integrated. Though hundreds of species of Green River fish, insects, and land plants had been described, the unusual microscopic aquatic flora and fauna were practically unknown before his studies were undertaken. As described in his 1931 publication, these fossils include a great array of bacteria, spores and fragmentary vegetative parts of fungi, algae, spores and pollen of higher plants, rhizopods, insects, and the earliest American mite. Among the rhizopods is the earliest known thecamoebian (*Difflugia calcifera*). The Green River also contains many unusual authigenic minerals, particularly saline ones, five of which are unknown elsewhere and two of which are known elsewhere only in igneous or metamorphic rocks (Milton and Eugster, 1959, p. 124). Bradley himself described the occurrence of some of the minerals. His most notable publication on the Green River formation, however, is "The varves and climate of the Green River epoch" (1929), a classic in a little touched field: the determination of the duration in years of a pre-Pleistocene epoch by means of a penetrating analysis of a purely sedimentary record. By demonstrating that pairs of laminae, one of which is richer in organic matter than the other, are to be interpreted as annual deposits, an estimate of about 6 million years was reached for the duration of Green River time—the longest span of time yet dated by a sedimentary record. That, combined with his estimate of the rate of deposition

of the underlying Wasatch and the overlying Bridger and Uinta formations (all fluviatile), indicates a little less than 23 million years for the Eocene epoch.

Bradley's skill as a draftsman and artist—an avocation from which both he and Mrs. Bradley draw much satisfaction—is shown by illustrations in his "Geomorphology of the north flank of the Uinta Mountains" (1936). This study was a byproduct of the Green River work. Its purpose was to integrate the early Tertiary depositional history of the Green River basin, north of the Uinta Mountains, with the later erosional history.

The latest published byproduct of the Green River investigations was a spectacular venture in paleoscatology, "Coprolites from the Bridger formation of Wyoming; their composition and microorganisms" (1946), which was completed as the United States entered World War II. One of the mammalian coprolites, inferred to be the product of a canivore or omnivore, in particular contains well preserved microscopic fossils: bacteria and algae, chiefly minute desmids. It was concluded that the animal had a last drink from a pool of acid swamp water. The chemical analysis of the coprolite is by Bradley himself, a token of his early training in chemistry.

Bradley was a submarine geologist before that term came into general use. "Geology and biology of North Atlantic deep-sea cores between Newfoundland and Ireland" (1941-42) records a wealth of data, deductions, and speculations on pioneer deep-sea cores from the North Atlantic Ocean. In addition to organizing the investigation, Bradley collaborated with M. N. Bramlette in the preparation of the chapter on lithology and geological interpretations. Four zones characterized by an abundance of sand and subrounded to angular pebbles were interpreted as glacial marine deposits formed by debris from drift ice, probably during minor advances of the last main glaciation. The cores between these zones consist almost entirely of foraminiferal marl. Two zones of silicic volcanic ash aided in correlations. The suggestion that layers sharply set off by regular upward gradation from coarse to fine (graded bedding) are the result of submarine slumping and deposition beyond the slide itself of material thrown into suspension was prophetic. It was the first time that mode of transportation and deposition was invoked for deep-sea sediments.

As Chief of the Military Geology Unit, Bradley recruited and directed a staff of almost 100 geologists, engineering geologists, geomorphologists, groundwater geologists, soils scientists, botanists, cartographers, and translators. The Unit prepared for the Army Corps of Engineers reports and maps on terrain intelligence, trafficability, water supply, sites for airfields and beach landings, and other matters involving terrain and geology. Most of the work was done under demanding urgency. Some of it had the flavor of mystery stories, as when a paleontologist on the staff (K. E. Lohman) was able, through the identification of diatoms recovered from ballast sand, to point to the launching sites of the small Japanese balloons that reached the Pacific coast. The sites were confirmed at the close of the war.

Bradley will long be remembered for his 15-year tour of duty as Chief Geologist. When he came to Washington in 1922, the Geologic Division had some 120 geologists and other professional employees, but when he became Chief Geologist that number had risen to some 400 and it reached a maximum of almost 1,000 in 1955. It was his responsibility to plan and direct, through his Branch Chiefs, the research activities of that large staff. During his tenure of office new vistas in geology were unfolding. He had the foresight to see and anticipate them and the good judgment to explore and expand them. His greatest satisfaction came from seeing new frontiers occupied and bearing fruit, his greatest frustration from the realization that his plans could not be fulfilled on account of the lack of personnel or funds. He inspired loyalty, not only in his Assistant Chief Geologists and Branch Chiefs, with whom he was in closest contact, but throughout the Geologic Division. He enjoyed dealing with people and his relations with them were tempered with his innate modesty and the milk of human kindness, although he could be tough when the rare occasion demanded it.

He could not have borne the exacting demands of his responsibility without respite through contact with the earth. Several summers were spent in the west in the continuation of work on the Green River formation, and others in Maine. The field work in Maine was part of a program of investigations in an attempt to throw light on the declining population of soft-shelled clams (*Mya arenaria*) in some of the Maine bays. Bradley (1957) has published the results of an exhaustive study of the physical and ecologic features of the Sagadahoc Bay tidal flat. His most recent publication (1959), a byproduct of the Sagadahoc Bay study, deals with a charming little clam that has the alliterative name *Gemma gemma*. This little clam and *Mya* tend to be incompatible and evidently compete with each other, either for food or for some other reason.

Bradley's achievements have brought him honors. He is a member of the National Academy of Sciences and the American Academy of Arts and Sciences. In 1946 he was President of the Geological Society of Washington and in 1953 a Vice-President of the Geological Society of America. The Washington Academy of Sciences gave him its Award in the Physical Sciences in 1940. On the occasion of the centennial of the Sheffield Scientific School of Yale University in 1947, his alma mater awarded him the degree of Doctor of Science. In 1958 the Interior Department honored him with its highest award, the Distinguished Service Award and Medal. The Guidebook of the Geology of the Uinta Basin prepared for the Eighth Annual Field Conference of the Intermountain Association of Petroleum Geologists in 1957, was dedicated to him. He has been an Associate Editor of this Journal since 1945.

After an interruption of 17 years, he is at present engaged in the study of the Green River formation of the Bridger, Washakie, and Green River basins —a study all the more exciting on account of recent advances in geochemistry.

W. P. WOODRING

PRINCIPAL PUBLICATIONS OF WILMOT HYDE BRADLEY

1924, Fossil caddice fly cases from the Green River formation of Wyoming: AM. JOUR. SCI., ser. 5, v. 7, p. 310-312.
1924, An oil shale and its microorganisms from the Fuson formation of Wyoming: AM. JOUR. SCI., ser. 5, v. 8, p. 228-234.
(With Sears, J. D.) 1924, Relations of the Wasatch and Green River formations in northwestern Colorado and southern Wyoming: U. S. Geol. Survey Prof. Paper 132, p. 93-107.
1925, A contribution to the origin of the Green River formation and its oil shale: Am. Assoc. Petroleum Geologists Bull., v. 9, no. 2, p. 247-262, 410.
1926, Shore phases of the Green River formation in northern Sweetwater County, Wyoming: U. S. Geol. Prof. Paper 140, p. 121-131.
1929, The occurrence and origin of analcite and meerschaum beds in the Green River formation of Utah, Colorado, and Wyoming: U. S. Geol. Survey Prof. Paper 158, p. 1-7.
1929, The varves and climate of the Green River epoch: U. S. Geol. Survey Prof. Paper 158, p. 87-110.
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1929, Neue Beobachtungen uber Algen als Urmaterialien der Bogheadkohlen und schiefer: Centralblatt f. Min., Abt. B. No. 5. p. 182-190.
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(and Pepper, James Franklin) 1938, Structure and gas possibilities of the Oriskany sandstone in Steuben, Yates, and parts of the adjacent counties, Pt. 1 *of* Geologic structure and occurrence of gas in part of southwestern New York: U. S. Geol. Survey Bull. 899-A, iv, 68 p.
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1940, Geology and climatology from the ocean abyss: Sci. Monthly, v. 50, no. 2, p. 97-109.
1940, Pediments and pedestals in miniature: Jour. Geomorphology, v. 3, no. 3, p. 244-254.
1945, U. S. Geological Survey Oil and gas investigation preliminary map no. 32, Geology of the Washakie Basin, Sweetwater and Carbon counties, Wyoming, and Moffat county, Colorado.
1946, Coprolites from the Bridger formation of Wyoming, their composition and microorganisms: AM. JOUR. SCI., v. 244, p. 215-239.
1947, A suggested geological curriculum: Geol. Soc. America Interim Proc., Pt. 2, p. 8-13.
1948, Limnology and the Eocene lakes of the Rocky Mountain region (U.S.): Geol. Soc. America Bull., v. 59, no. 7, p. 635-648.
1957, Physical and ecologic features of the Sagadahoc Bay tidal flat, Georgetown, Maine: Geol. Soc. America Mem. 67, p. 641-682.
(with Peter Cooke) 1959, Living and ancient populations of the clam *Gemma gemma* in a Maine coast tidal flat: U. S. Fish and Wildlife Service, Fishery Bull. 137, p. 305-334.

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MAGNETIC SUSCEPTIBILITY ANISOTROPY AND FABRIC OF SOME ADIRONDACK GRANITES AND ORTHOGNEISSES*

J. R. BALSLEY[1] and A. F. BUDDINGTON[2]

ABSTRACT. The magnetic susceptibility anisotropy of about 150 specimens of granites and orthogneisses of the northwest Adirondack area has been determined. The maximum susceptibility of the orthogneisses is uniformly about parallel to the lineation, and the minimum susceptibility is characteristically about at right angles to the planar structure. A high ratio of maximum to intermediate magnetic susceptibility indicates a high intensity of linear structure. A high ratio of the mean of the maximum and intermediate to the minimum magnetic susceptibility indicates a high intensity of planar structure. Studies of magnetic susceptibility anisotropy have permitted the determination of lineation in many rocks where it could not be determined in the field and have afforded quantitative bases for correlations with broad geologic structures.

INTRODUCTION

This paper presents the results of a study of the magnetic susceptibility anisotropy of some Precambrian granites and orthogneisses of the northwest Adirondack area in New York state. It is part of a continuing series of studies by the authors on the origin and character of the magnetic and paleomagnetic properties of these rocks.

Balsley is responsible for the geophysical aspects of the problem and Buddington for the geology and sample collection. We are indebted to William Huff of the U. S. Geological Survey for the actual determination of the magnetic data.

ROCK FABRICS

Three general associations of rocks (Buddington, 1943) in the northwest Adirondacks form the principal basis for this study: (1) a group of pyroxenic syenitic to hornblendic granitic orthogneisses of the Diana, Stark and Tupper complexes that are ordinarily thoroughly deformed and recrystallized, (2) a younger group of pink hornblende to biotite granitic rocks that range from gneissoid mesoperthite granites to wholly recrystallized pink to gray oligoclase-microcline granite gneiss and (3) an almost massive green fayalite-ferrohedenbergite mesoperthite granite that is the youngest granitic intrusive and is wholly unmetamorphosed. The term mesoperthite is used for a microperthite feldspar that consists of about equal parts of potassium feldspar and plagioclase. On metamorphism and recrystallization a mesoperthite feldspar may break down

* Publication authorized by the Director, U. S. Geological Survey.

[1] U. S. Geological Survey, Washington, D. C.

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or unmix into an aggregate of separate potassium feldspar and plagioclase grains.

Planar structure or foliation characterizes most of the rocks. It has had two modes of origin. One is a magmatic flow structure formed as the result of preferred orientation of inequidimensional crystals in the early stages of crystallization and subsequent overgrowth, locally succeeded in late stages by some crushing of already crystallized material but before complete consolidation of the mass. Planar structure of this type will be called *gneissoid*. The second type is the result of deformation in the solid state with plastic flowage at such temperatures that it is accompanied by recrystallization of the minerals. This will be referred to as *tectonoplastic flow* and the resulting rock is a *gneiss*. Such metamorphism results in a rock with a generally granoblastic texture in which the feldspars occur as a mosaic of polygonal grains and the quartz as elongate flattened lenses. The magnetite occurs as inequidimensional grains or in aggregates or trains of grains with a preferred orientation parallel to a linear structure in the gneissic rocks, and in inequidimensional grains with a preferred orientation in the gneissoid rocks.

The megascopic lineation as observed in the field is the result of a preferred dimensional orientation of inequidimensional mineral grains or mineral aggregates and invariably lies in the plane of foliation. A general discussion of the relations of lineation to other structures in Adirondack rocks has been previously published (Buddington, 1956).

The term "a" lineation will be used for a mineral lineation that is within 30° of being at right angles to the predominant trend of the formations and the major fold axes and subparallel to subordinate crossfolds, and "b" lineation for a mineral lineation that is within 30° of being parallel to the predominant trend of the formations and the major fold axes.

MAGNETIC SUSCEPTIBILITY ANISOTROPY

Graham (1954, p. 1257-1258) has pointed out that

> The susceptibilities in all directions of a sample with many elongate ferromagnetic grains disseminated in a neutral groundmass can be represented by an ellipsoid whose greatest axis parallels the preferred orientation of the long dimensions of the ferromagnetic grains.

S. W. Stewart, for a bachelor's thesis at Princeton University, studied 23 cores of northwest Adirondack gneisses and concluded that the direction of maximum magnetic susceptibility lies in the plane of foliation and is parallel to the megascopic lineation.

A detailed discussion of the iron and iron-titanium oxide mineralogy of the rocks has previously been published (Balsley and Buddington, 1958). The

mineral controlling the magnetic susceptibility of the rocks studied here, however, is magnetite.

The magnetic susceptibility of the rocks was measured with a 280 cycle Maxwell inductance bridge almost identical in design to that built by J. W. Graham at the Department of Terrestrial Magnetism of the Carnegie Institution of Washington. All components of the six-leg bridge, (shown diagrammatically

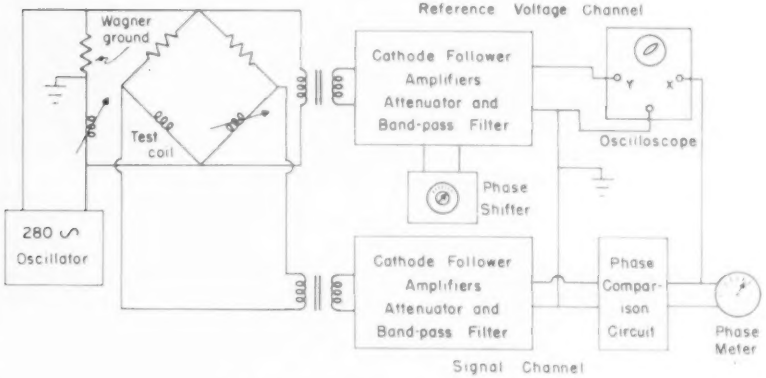

Fig. 1. Magnetic susceptibility bridge.

in fig. 1), are identical to reduce the effect of temperature variation. The inductive legs are identical coils wound on the Helmholtz principle so that the field in the volume occupied by the sample is homogenous to within one percent. The output of the oscillator is adjusted so that a magnetic field of ± one gauss exists within the coils. The inductance of the Wagner ground leg of the bridge is changed by moving a single turn of the manganin wire for the resistive balance and by moving a ferrite-permalloy slug for the inductive balance. The phase angle between the reference voltage and the signal is adjusted so that the phase comparison circuit is insensitive to the resistive component. By this means, differences of more than an order of magnitude in the resistivities of sample and standard do not disturb the inductive balance. After the bridge has been balanced at zero, the sample is inserted and the ferrite-permalloy slug of the measuring coil is moved on a screw until the bridge is rebalanced. The position of the slug is determined by the screw that has been calibrated empirically to give readings of the magnetic susceptibility.

Ferric chloride solution was used for the calibration in the lower range of susceptibility and in the middle range, samples measured on the bridge of

Department of Terrestrial Magnetism were used. In the higher range, samples of magnetite were used whose susceptibility was computed from observations of the anomalous field they had produced as measured on the equipment of the Naval Ordnance Laboratory. Because there are no standards of magnetic susceptibility above 10^{-5} cgs per cc with low conductivity, the absolute accuracy of the susceptibility measurements on this equipment cannot be determined. The consistency of all three overlapping calibrations discussed above, however, indicates that the accuracy is on the order of one percent for susceptibilities between 10^{-6} and 0.5 cgs per cc.

The susceptibility anisotropy is measured by inserting within the test coil a non-magnetic holder modeled after a three axis universal stage. The sample is rotated about two axes to seek out either the maximum or minimum susceptibility. The direction and intensity of this is recorded and the sample is rotated 90° about the horizontal axis that is at right angles to the axis of the coil, also horizontal. The sample is then rotated about the vertical axis to determine the direction and intensity of the other two susceptibility axes. This system, of course, assumes that the first measured axis is at right angles to the other two, but in all rocks we have measured, the angular separation between the last two measured axes has been 90°, within the limit of measurement. Because it is necessary to construct the universal stage holder of non-conductive and non-inductive materials and because space is limited within the coil, the precision of the angular measurement of the susceptibility is no better than ± 5°.

Data for the discussion of possible systematic relationships between magnetic susceptibility and anisotropy and rock fabrics such as planar foliation and lineation are given in Table 1.

INDISTINCTLY GNEISSOID FAYALITE-FERROHEDENBERGITE MESOPERTHITE GRANITE

A small arcuate body of green fayalite-ferrohedenbergite granite occurs north of Wanakena (fig. 2) (Cranberry Lake quadrangle). It is about 9 miles long and ½ to one mile wide. The rock is in general so massive that planar structure can be discerned only with difficulty and lineation not at all. The granite is wholly unmetamorphosed. There is, accordingly (table 1, belt 1), the least spread (1.20:1.13:1) in the relative intensities of the maximum, intermediate, and minimum susceptibilities for any of the Adirondack rocks studied. Furthermore the directions of the susceptibility axes of different samples are unsystematic.

GNEISSOID HORNBLENDE AND BIOTITE MESOPERTHITE GRANITES

Part of a great elongate batholith of hornblende and biotite mesoperthite granite extends northeast from Lowville and Glenfield (fig. 2) through Kirschnerville. It is younger than the orthogneisses of the Diana, Stark, and Tupper complexes.

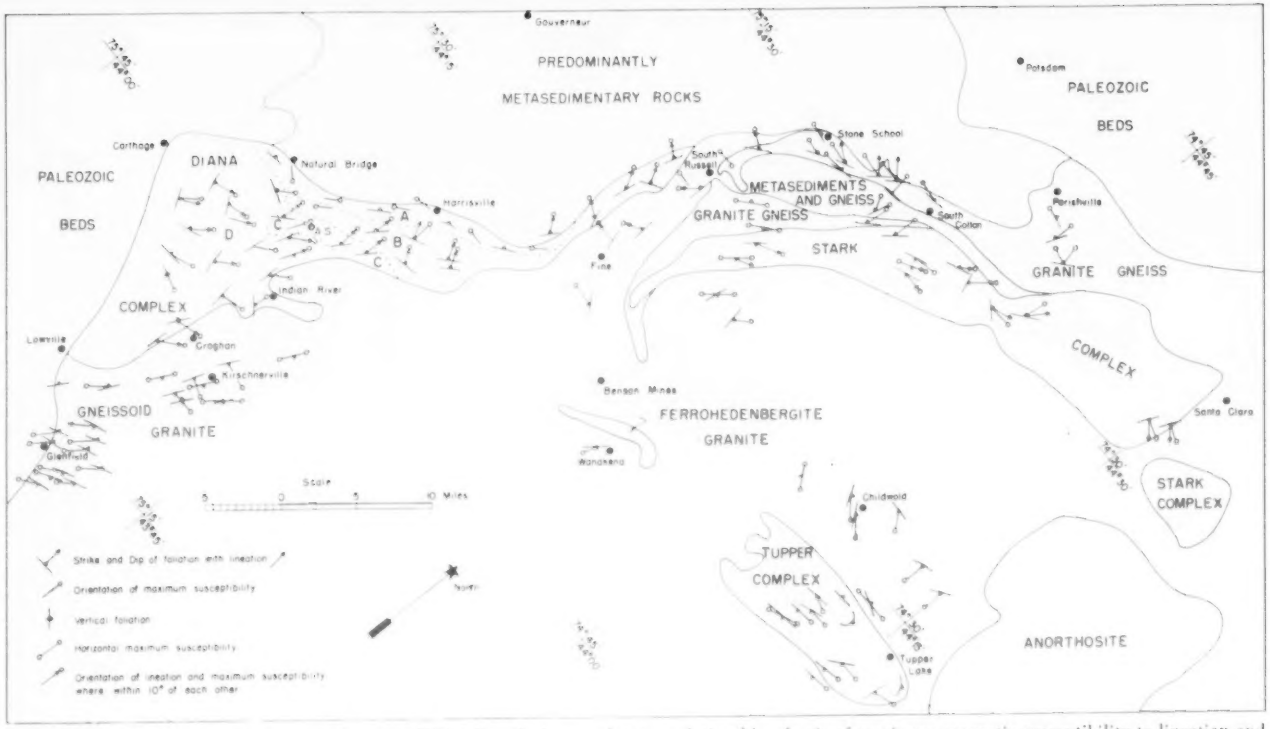

Fig. 2. Simplified geologic map of part of NW Adirondack area showing relationship of axis of maximum magnetic susceptibility to lineation and other structures. Diana Complex: A—Belt of mortar gneiss without discernible lineation maximum susceptibility sub-parallel to planar foliation. B—Belt of recrystallized granoblastic gneiss with maximum susceptibility parallel to strong "a" north-northwest lineation. C—Belt of granoblastic gneiss with maximum susceptibility parallel to north lineation, in part "b" lineation, in part transitional to "a" lineation. D—Belt of recrystallized generally granoblastic gneiss with maximum susceptibility parallel to strong east-northeast to north-northeast lineation. AS—Aldrich School.

TABLE 1

Data on magnetic susceptibility anisotropy and related rock fabric

Belts **	No. of specimens	Magnetic Measurements: Susceptibility: Axis	Az.	Incl.	Int.*	Int.	Ratio: Max. & Inter. to 2 Min.	Ratio: Max. to Inter.	Field Observations: Foliation: Az.	Foliation: Dip	Foliation: Descrip.	Lineation: Az.	Lineation: Descrip.
		Max			0.75	1.20							
1	7	Interm	Unsystematic		0.71	1.13	1.16	1.06		Indistinct			None
		Min			0.63	1.00							
			7 - 11°	3°									
		Max	6 - 92°	61°	1.21	1.35							
			3 - 0°	90°									
			4 - 270°	43°									
2	16	Interm	3 - 118°	26°	1.10	1.22	1.29	1.10	25°	70°E	Good Gneissoid		Indiscernible
			9 - 191°	10°									
			8 - 291°	5-25°; 3-52°									
		Min	3 - 90°	47°	0.90	1.00							
			3 - 124°	0°									
		Max	43°	4°	1.60	1.32							
2-a	4	Interm	310°	45°	1.40	1.15	1.24	1.14	29°	50°W	"		Indiscernible
		Min	137°	45°	1.21	1.00							
		Max	234°	5°	2.61	1.58							
3	9	Interm	Unsystematic		2.04	1.24	1.40	1.28	60°	60°NW	"	Gentle SW	Good to faint
		Min			1.65	1.00							
		Max	230°	10°	2.10	1.36							
3-a	14	Interm	131°	34°	1.82	1.18	1.28	1.15	58°	52°SE	"		Indiscernible
		Min	330°	48°	1.54	1.00							
		Max	133°	13°	5.94	1.30							
4	4	Interm	38°	28°	5.19	1.13	1.21	1.15	143°	56°NE	Good Gneissic		Indiscernible
		Min	226°	52°	4.59	1.00							
		Max	286°	17°	2.80	1.56							
5	5	Interm	Unsystematic	0°	2.51	1.42	1.49	1.12	40°	15°NW	"	292°	Moderate
		Min	122°	80°	1.79	1.00							

TABLE 1 (Continued)

Belts **	No. of specimens	Axis	Magnetic Measurements: Susceptibility Az.	Incl.	Int.*	Int.	Ratio: Max. & Inter. to 2 Min.	Max. to Inter.	Field Observations: Foliation Az.	Dip	Descrip.	Lineation Az.	Descrip.
		Max	280°	13°	3.42	1.45							
5-a	4	Interm	188°	0°	2.78	1.17	1.31	1.23	70°	15°N	"	290°	Good
		Min	105°	83°	2.36	1.00							
		Max	275°	18°	3.43	1.33	1.28						
5-b	4	Interm	184°	0°	3.16	1.22		1.09	68°	26°N	Good Gneissic	290°	Indistinct
		Min	?	70°	2.59	1.00							
		Max	279°	31°	3.04	1.30							
5-c	6	Interm	?	23°	2.93	1.26	1.30	1.04	62°	50°NW	"		Indiscernible
		Min	141°	44°	2.33	1.00							
		Max	43°	0°	2.55	1.54							
5-d	2	Interm	—	—	2.30	1.40	1.46	1.11	40°	30°NW	"		Indiscernible
		Min	120°	75°	1.66	1.00							
		Max	—	4° - 55°	5.35	1.50							
6	9	Interm	—	27°	4.81	1.35	1.40	1.11	30° - 65°	60°NW	Good Gneissoid		Indiscernible
		Min	154°	39°	3.64	1.00							
		Max	337°	34°	4.67	2.15							
6-a	14	Interm	242°	15°	3.83	1.77	1.96	1.22	47°	40°NW	Intense Gneissic	327°	Intense "a"
		Min	130°	53°	2.17	1.00							
		Max	13°	42°	3.78	1.91							
6-b	5	Interm	275°	34°	3.34	1.70	1.81	1.13	39°	52°NW	Strong Gneissic	5°	Strong
		Min	143°	40°	1.97	1.00							
		Max	20°	Variable	3.39	1.42							
6-c	3	Interm	269°	31°	2.75	1.15	1.29	1.23	20°	Variable	Indistinct Gneissic	20°	Intense "b" on nose of fold
		Min	125°	Variable	2.39	1.00							
		Max	45°	35°	3.25	1.77							
6-d	13	Interm	8 = 312° 4 = 173°	21°	2.44	1.33	1.55	1.33	338°	35°E	Good Gneissic	59°	"
		Min	222°	52°	1.84	1.00							

TABLE 1 (Continued)

Belts **	No. of specimens	Magnetic Measurements: Susceptibility Axis	Az.	Incl.	Int.*	Int.	Ratio: Max. & Inter. to 2 Min.	Ratio: Max. to Inter.	Field Observations: Foliation Az.	Foliation Dip	Foliation Descrip.	Lineation Az.	Lineation Descrip.
		Max	**53°**	**26°**	**2.79**	1.40							
6-e	3	Interm	2 = 236°	57°	2.12	1.07	1.23	1.32	73°	61°N	Poor Gneissic	57°	"
		Min	316°	0°	**1.99**	1.00							
		Max	**74°**	**19°**	**2.40**	1.46							
6-f	4	Interm	Variable	Variable	1.95	1.18	1.32	1.23	324°	43°NE	Moderate Gneissic	75°	"
		Min	Variable	64°	1.65	1.00							
		Max	45°	0°	**1.92**	**1.47**							
7	3	Interm	132°	53°	1.69	1.29	1.37	1.14	40°	52°SE	Moderate Gneissic	60°	Slight-Indiscernible
		Min	314°	37°	1.31	1.00							
		Max	80°	10°	1.02	1.31							
7-a	2	Interm	166°	**48°**	0.91	1.17	1.24	1.12	65°	52°SE	"		Indiscernible
		Min	334°	37°	0.78	1.00							
		Max	81°	23°	2.11	1.33							
7-b	5	**Interm**	188°	50°	1.90	1.20	1.28	1.11	66°	43°SE	"		"
		Min	338°	28°	1.58	1.00							
		Max	47°	0°	**1.77**	1.51							
7-c	2	Interm	137°	30°	1.37	1.17	1.34	1.29	340°	Gentle E	"	50°	Intense "b" on nose of fold
		Min	317°	60°	1.17	1.00							
		Max	109°	29°	1.29	1.66							
7-d	3	Interm	16°	15°	1.06	1.36	1.51	1.22	28°	38°E	**Good** Gneissic	125°	"
		Min	287°	51°	0.78	1.00							
		Max	82°	10°	1.85	1.70							
8	10	Interm	Unsystematic	6°	1.57	1.46	1.58	1.18	74°	30°S	"	75°	Indiscernible-faint "b"
		Min		73°	**1.08**	1.00							
		Max			2.03	1.26							
9		Interm	Unsystematic		1.82	1.13	1.20	1.12	N	Steep	Layered Ore		Indiscernible
		Min	96°	15°	1.61	1.00							

* in units of 10^{-3}cgc/cc

**

1) Fayalite-ferrohedenbergite mesoperthite granite; north of Wanakena.
2) Hornblende and biotite mesoperthite granite, belt through Kirschnerville.
2-a) Same as (2), 7 miles southwest of Kirschnerville.
3) Hornblende and biotite microperthite granite; northwest limb of Glenfield anticline.
3-a) Same as (3) but on southeast limb of Glenfield anticline.
4) Hornblende microcline-rich granite gneiss with pegmatite seams and amphibolitic schliere, south and west of Childwold.
5) Hornblende microcline-oligoclase granite gneiss, 4-6 miles SW of Stone School.
5-a) Same gneiss as (5); Stone School.
5-b) Same gneiss as (5); 2 miles east of Stone School.
5-c) Same gneiss as (5); 5-8 miles ENE of Stone School, about 4 miles W of South Colton.
5-d) Same gneiss as (5); 3.5 miles NE of South Russell.
6) Green pyroxene quartz syenite mortar gneiss, border facies "A", (fig. 2) of Diana complex.
6-a) Pyroxene or hornblende quartz syenite gneiss; northwest lineation; Diana complex, belt "B".
6-b) Hornblende or pyroxene quartz syenite gneiss; Diana complex, belt "C".
6-c) Pyroxene quartz syenite gneiss; Diana complex, belt "C".
6-d) Pyroxene or hornblende quartz syenite gneiss; Diana complex, belt "D".
6-e) Hornblende or pyroxene granite gneiss; Diana complex, belt "D".
6-f) Similar gneisses to 6-e, Diana complex, belt "D".
7) Hornblende or hornblende-pyroxene granite gneiss; Stark complex; about 8 miles SSE of Stone School.
7-a) Hornblende granite gneiss; Stark complex, about 10.5 miles east of Stone School.
7-b) Hornblende and pyroxene quartz syenite gneisses; Stark complex, near French Lake about 3.25 miles SSE of South Colton.
7-c) Green hornblende-pyroxene quartz syenite gneiss, 3 miles ESE of South Colton.
7-d) Hornblende and pyroxene granite gneiss; on east-southeast plunging nose of anticline, Stark complex; about 5 miles south of Santa Clara.
8) Pyroxene quartz syenite (charnockite) gneiss; Tupper complex.
9) Low grade magnetite ore, Newton Falls pit, Benson Mines.

A gneissoid facies of the granite is well exposed to the south and west of Kirschnerville. The rock ranges from a hornblende granite to a biotite alaskite. The rocks are effectively unmetamorphosed and the mesoperthite shows only a little segregation of exsolved plagioclase at the borders of the grains. There is a well developed gneissoid flow structure that dips steeply. Lineation can rarely be observed.

In the granite of the Kirschnerville belt with gneissoid foliation having an easterly dip (table 1, belt 2) the axes of maximum magnetic susceptibility lie close to the plane of foliation either sub-horizontal or sub-parallel to the direction of maximum dip. Some of the axes of intermediate susceptibility are sub-horizontal and close to the plane of foliation but some are not. Some of the axes of minimum susceptibility are sub-perpendicular to the plane of foliation but some are not. Inspection of polished surfaces of the granites shows that the orientation of the axes of inequidimensional grains of magnetite relative to the plane of foliation may vary widely. Some of the long and intermediate axes of the inequidimensional grains lie in the plane of foliation but some of the long axes of the grains may be at right angles to or oblique to the plane of foliation. It is probable that this varied orientation of the inequidimensional grains of magnetite gives rise to the diverse orientation of the susceptibility axes relative to the plane of foliation. More detailed studies would be necessary to determine the origin of the apparently abnormal relationships of susceptibility

axes to the plane of foliation.

In one area of granite (table 1, belt 2-a) of the Kirschnerville belt where the foliation dips *west*, the axes of the maximum and intermediate susceptibility lie about in the plane of foliation and the axes of minimum susceptibility are about at right angles to the planar structure.

The granite of the batholith near Glenfield is well exposed on both the east and west sides of the Black River valley and has an anticlinal structure that is called the Glenfield anticline. The rock, as in the Kirschnerville belt ranges from a hornblende granite to a biotite alaskite. It has a well developed planar structure, effectively from magmatic flow, but the lineation though present locally is generally indistinct. It is inferred to be a "b" lineation. The planar structures indicate an anticline plunging gently (30° ±) southwest. Dips in the axial zone are gentle but in the limbs are generally 60° or more. Microscopic study shows that the gneissoid structure is due to preferred orientation of: inequidimensional feldspars, hornblende and biotite; elongate amoeboid quartz; and local leaves of finer grained feldspar and quartz aggregates. The feldspar of the granulated aggregates is largely mesoperthite but it is in part unmixed to plagioclase and potash feldspar. The feldspar grains of the granulated material have irregular borders that mutually interlock with the primary grains. The texture of the granite is thus in large part primary magmatic but to a small extent developed by movement, perhaps just before complete consolidation of the mass.

The granite of the Glenfield anticline (table 1, belts 3 and 3-a) uniformly has the axes of maximum of magnetic susceptibility about parallel to the strike of the foliation, usually with a gentle plunge to the southwest. This is consistent with the structure of the anticline. In the northwest limb the axes of intermediate and minimum of susceptibility are unsystematically related to the foliation. The slightly more gneissoid granite (table 1, belt 3-a) of the southeast limb, however, has all the axes of susceptibility in systematic relationship to the foliation. The intermediate axes are close to the plane of foliation and the axes of minimum susceptibility are at right angles to the plane of foliation.

The ratio of the mean of the intensities of the maximum and intermediate susceptibilities to the intensity of minimum magnetic susceptibility in the granites and gneiss generally gives an idea of the intensity of development of planar structure. The ratio for the fayalite-ferrohedenbergite granite that has only indistinct planar structure is 1.16, for gneissoid granite of the Kirschnerville belt and Glenfield anticline with good planar structure it is 1.24 to 1.40. Linear structure is only observable in the gneissoid granites in the field where the ratio of maximum to intermediate susceptibility is greater than about 1.2.

PEGMATITE-SEAMED CONTAMINATED HORNBLENDE GRANITE GNEISS

West of Childwold (fig. 2) there is a belt of pegmatite-seamed contaminated hornblende granite gneiss with included layers and schliere of amphibolite. The pegmatite seams are thin and relatively massive; the gneissic part is a microcline-rich facies of granite gneiss. The abnormally large susceptibility is due to a relatively high content of magnetite, common in contaminated facies of granitic rocks. In the field the alternating pegmatite seams and

gneissic granite give to the rock as a whole a conspicuous layered or planar structure. The massiveness of the pegmatitic facies in the rock, however, gives rise to a relatively low ratio (table 1, belt 4) of the mean of the maximum and intermediate susceptibility intensities to that of minimum susceptibility. No linear structure could be observed in the field. The axes of the maximum susceptibility are in the plane of foliation, the axes of intermediate susceptibility lie close to it and the axes of minimum susceptibility are about at right angles to the planar structure.

HORNBLENDE GRANITE ORTHOGNEISS, STONE SCHOOL BELT

An arc of hornblende granite orthogneiss extends east northeast and south through Stone School (fig. 2). This belt of gneiss forms the northwest limb and southwest end of an overturned isoclinal syncline that plunges gently northeast to east-northeast. The northwest limb dips gently, 20°-30° northwest. There is a well developed tectonoplastic flow planar structure. A well developed "a" lineation strikes generally west-northwest and is parallel to the axis of a crossfold that forms an embayment toward Parishville. Microscopic study shows the feldspar of the gneiss to consist of a granoblastic aggregate of about equal parts of oligoclase and a slightly microperthitic microcline. The grains of the feldspars averages 0.6 to 1.2 mm in diameter. The hornblende is a blueish-green variety. The opaque oxide is exclusively magnetite which contains not more than 2 percent TiO_2. Sphene is an accessory mineral. These characteristics indicate that the gneiss was metamorphosed in the amphibolite facies at a temperature estimated between 450°-550°C.

The magnetic susceptibilities are interesting in that the spread between the maximum and minimum is less where the rock strikes N.65°E (table 1, belts 5-b, 5-c) than where it strikes N 40°E (table 1, belt 5). This accords with the fact that in general the formations trend about northeast and the maximum deformation might be expected in rocks with this strike. The azimuth of the lineation and of the maximum susceptibility remains the same in rocks of differing strike.

The magnetic susceptibility anisotropy correlates with a systematic change from a gneiss with strongly developed planar and weak linear structure (table 1, belt 5) 4-6 miles southwest of Stone School through a rock with moderate planar and linear structure (table 1, belt 5-a) near Stone School to a gneiss several miles east of Stone School with weak planar and at best indistinct linear structure (table 1, belts 5-b, 5-c).

The axes of maximum susceptibility lie in or close to the plane of foliation subparallel to the lineation and the axes of minimum susceptibility are, within 10 degrees or less, at right angles to the foliation.

DIANA COMPLEX OF GRANITIC AND SYENITIC ORTHOGNEISS

The Diana complex (Buddington, 1948, p. 24-30) has been interpreted as a differentiated thick sheet of igneous rock that has been intensely folded and metamorphosed. It varies from green pyroxene syenite gneiss with inter-

layers or lenses of shonkinite gneiss through pyroxene-hornblende and pink hornblende quartz syenite gneiss to hornblende granite gneiss. The sheet has a maximum stratigraphic thickness of about 19,000 feet southeast of Harrisville and thins to the northeast.

The Diana complex extends from Lowville through Harrisville to just west of South Russell (fig. 2).

The northwest border facies of the Diana complex for about ¼ to ½ mile width from Natural Bridge to northeast of Harrisville, is a coarse grained rock with a small amount of mortar structure. There is a preferred orientation of the inequidimensional minerals and some layering so that a planar structure is clearly defined. The layering and planar structure are primarily of magmatic origin and the rock has not been sufficiently deformed to develop a linear structure that can be discerned.

Southeast of the border facies and throughout most of the complex the rocks are wholly recrystallized to a granoblastic gneiss. The coarseness of grain increases slightly from northwest to southeast about at right angles to the foliation. The quartz is in flattened elongate lenses. The coarse primary mesoperthite grains of the original igneous rock have been recrystallized to small lenticular granoblastic aggregates or mosaics of polygonal grains of discrete microcline and plagioclase. In the southeast facies of the complex there are distinct porphyroblastic hornblendes in the quartz syenite and granite gneisses and porphyroblastic pyroxene in the pyroxene syenite gneiss. Relic porphyroclasts occur locally throughout the complex. The temperature of metamorphism is estimated as between 450°C and 550°C.

The general trend of the major folds within the complex is northeast. Northeast of Aldrich School the sheet forms the southeast limb of an anticline overturned to southeast with a general dip of 50° northwest. Between Carthage and Croghan the sheet is deformed in a series of isoclinal folds trending west-northwest and strongly overturned to the south-southwest with dips of 25°-40° to the north and north-northeast. These are inferred to be on the northeast plunging nose of a major northeast-plunging anticlinal complex. Two miles southwest of Aldrich School there is an anticlinal core with a steep axial plane striking about north.

The rocks of the simple overturned limb northeast of Aldrich School (AS, fig. 2) to west of South Russell have a strong northwest to north-northwest "a" lineation. The gneisses of the anticline south of Aldrich School have an intense north to north-northeast "b" lineation. The lineation of the gneisses of the Diana complex in the belt from Carthage to Croghan strikes northeast and is an "a" lineation if considered in its structural relation to the orientation of the axial planes of the folds in this zone. If, however, the lineation is considered in relation to the predominant northeast trend of folds in general in the northwest Adirondacks it would be a "b" lineation.

In all the *gneissic* rocks the axes of maximum susceptibility are, within the limits of accuracy of our measurements, parallel with the lineations. This

relationship is peculiarly emphasized because it holds for both "a" and "b" types of lineation whose azimuths range from 82° to 292° between Lowville and Harrisville through an arc of about 150°.

In all the gneissic rocks of the Diana complex and in the granite gneiss of the Stone School belt the axes of intermediate susceptibility are, within the limits of measurement, in the plane of foliation and at right angles to the lineation. The axes of the minimum susceptibility in the gneissic rocks of the Diana complex are about at right angles to the plane of foliation.

Where the lineation is northwest to north (table 1, belts 6-a and 6-b) the rocks usually have both an intense planar and an intense linear structure but where the lineation is north-northeast to east-northeast (belts 6-c, 6-d, 6-e, 6-f) the lineation is much more intensely developed than the planar structure. In the former rocks the ratio of the mean of the maximum and the intermediate to the minimum susceptibility is 1.80-1.96 and the ratio of the maximum to the intermediate is 1.13-1.22 whereas in the more pencil-like gneisses these ratios are 1.23-1.55 and 1.23-1.33 respectively.

STARK COMPLEX OF QUARTZ SYENITE AND GRANITE ORTHOGNEISS

The rocks of the Stark complex (table 1, belts 7 to 7-d, inclusive) consist of pyroxene and hornblende quartz syenite and granite orthogneiss. They can be inferred to represent a tight anticlinally folded sheet, in part isoclinal and overturned. They have a well developed planar structure of tectonoplastic origin. The main belt of the gneiss has a "b" lineation. The axes of the maximum susceptibility are closely parallel to that of the lineation and both lie in the plane of foliation. The axes of the minimum susceptibility are about at right angles to the plane of foliation and the axes of the intermediate susceptibility are in the plane of the foliation about perpendicular to the lineation. The gneiss south of Santa Clara (table 1, belt 7-d) is on the nose of an anticline plunging east-southeast and has an intense lineation almost perpendicular to the strike and in the plane of foliation.

PYROXENE-HORNBLENDE QUARTZ SYENITE GNEISS OF MT. ARAB ANTICLINE OF TUPPER COMPLEX

A belt of green pyroxene syenite orthogneiss and green pyroxene-hornblende quartz syenite orthogneiss occurs on the Mt. Arab anticline across the northeastern part of the Tupper Lake quadrangle west of the village of Tupper Lake (fig. 2). The gneisses have a well developed planar structure in the form of a broad anticline with gentle to moderately dipping limbs and an easterly-plunging nose that is well exposed south of Tupper Lake. Hypersthene is a major to predominant member of the mafic minerals present so the rocks are of charnockitic character. They have been almost completely metamorphosed and recrystallized in the granulite facies and in general have a granoblastic texture. Locally the gneisses contain a few relic porphyroclasts of mesoperthite. The potash feldspar has a microcline structure and a micro-

perthitic intergrowth of plagioclase. The temperature of metamorphism has been estimated at around 600°C, or perhaps higher.

The dark color of the rocks and poor exposures have prevented quantitative determination of the azimuth of the lineation of the specimens studied but qualitatively it is known to be a "b" lineation about parallel to the axis of the fold. The axes of the maximum susceptibility (table 1, belt 8) corresponds with this. The axes of the intermediate and minimum susceptibility however are unsystematic for the samples collected.

IRON OXIDE ORES

The ore deposit in the Newton Falls pit at Benson Mines is low grade and consists of alternating thin layers of magnetite and gneiss. The ore in part has a flat network or mesh structure relative to the gneiss. The magnetite replaces the gneiss and is post-metamorphic. The axis of minimum magnetic susceptibility is about at right angles to the layering which strikes about due north and dips very steeply. The azimuths and angles of inclination of the maximum and intermediate susceptibilities are widely scattered however.

The Clifton ore deposit is a high grade magnetite replacement of skarn. Again the azimuth of the minimum susceptibility is at right angles to the strike of the ore body but the angle of inclination is not perpendicular to the dip of the ore. Also the azimuth of the maximum susceptibility is at substantial angles to the strike of the ore body.

In the *gneissoid* granites, with planar structure as good as some of that in the *gneissic* rocks, however, the relationships of the susceptibility axes to rock fabric may not be as systematically simple. In the gneissoid rocks the axes of the maximum susceptibility are generally parallel to the planar structure. The axes of minimum susceptibility though in substantial part at right angles to the plane of foliation may locally lie close to it. In part, the relationships of the susceptibility axes to rock fabric of the *gneissoid* granites may be similar to and as systematic as those in the *gneissic* rocks, but in the least gneissoid granite the orientation of the susceptibility axes varies greatly.

SUMMARY AND CONCLUSIONS

A high ratio of maximum to intermediate susceptibility indicates a high intensity of linear structure and a high ratio of the mean of the maximum and intermediate susceptibilities to the minimum susceptibility indicates an intense development of the planar structure or flattening.

The most intense linear structure is found in general in two quite different structural relationships. One mode of occurrence is the belt of orthogneiss of the Diana complex (table 1, belt 6-a) where an "a" type of lineation prevails and the rock also has an intense planar structure. The other mode of occurrence is where a "b" lineation occurs on the noses of gently to moderately plunging folds (table 1, belts 6-c, 6-d, 6-e, 6-f, 7-c, 7-d), and the planar structure may vary from indistinct to good.

In all the *gneissic* rocks the axes of maximum magnetic susceptibility are about parallel to the lineation and together with the axes of intermediate susceptibility lie in the plane of foliation. The axes of minimum susceptibility are usually about at right angles to the plane of foliation.

The hue and color contrast of the constituent minerals of the rocks is the major factor that permits clear identification of any lineation in the field. It was found impossible to distinguish the lineation, except very locally, in the green gneisses of the Stark and Tupper complexes although the susceptibility ratios (table 1, belts 7-a, 7-b, 7-c and 8) indicate a definite lineation whose azimuth is consistent with that expectable and with such field determinations as could be made. Where "b" lineation is intense on the nose of folds it may be so pronounced (table 1, belts 6-c, 6-e, and 7-c) that the gneissic planar structure can only be determined in the field with the utmost difficulty, if at all, though the susceptibility ratios are such as to suggest that a good gneissic planar structure would be present.

The magnetic susceptibility anisotropy studies have proved valuable in supplementing field and petrographic studies. They have permitted the determination of the lineation in many rocks where it could not be determined in the field and have afforded quantitative bases for correlations with geologic structures.

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THERMODYNAMIC PROPERTIES OF SOME SYNTHETIC ZINC AND COPPER MINERALS*

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ABSTRACT. Measurement of pH of a solution of known Zn^{++} or Cu^{++} concentration in equilibrium with the metal hydroxide or oxide allows the calculation of an apparent equilibrium constant ($K' = [M^{++}] \cdot a^2_{OH^-}$) which varies with composition of the solution. Extrapolation of K' values to infinite dilution permits estimation of a true equilibrium constant ($K = a_{M^{++}} \cdot a^2_{OH^-}$). This procedure is applied to the determination of the equilibrium constant for ZnO (zincite). $CuCl_2$ and $CuSO_4$ solutions precipitate the basic salts $Cu_4(OH)_6Cl_2$ (atacamite), and $Cu_4(OH)_6SO_4$ (brochantite) at concentrations greater than about $10^{-4.5}$ molar. Extrapolation to infinite dilution gives the equilibrium constants for these basic salts. At higher pH values reactions such as

$$Cu_4(OH)_6Cl_2 + 2OH^- \rightleftarrows 4CuO + 2Cl^- + 4H_2O \qquad (1)$$

and

$$Cu_4(OH)_6SO_4 + 2OH^- \rightleftarrows 4CuO + SO_4^{--} + 4H_2O \qquad (2)$$

take place and permit the estimation of K for CuO (tenorite). For zincite, $K = 10^{-17.0 \pm 0.1}$; for brochantite, $K = 10^{-20.6 \pm 0.4}$; for atacamite, $K = 10^{-20.4 \pm 0.4}$; and for tenorite, $K = 10^{-13.7 \pm 0.5}$. The variation of K with temperature (25° to 75° C) permits the calculation of heat content for tenorite, brochantite, and atacamite, but for tenorite the more reliable heat of formation is obtained from the free energy data and the entropy measured by Hu and Johnston (1953).

To the extent that the pH measurements represent the hydrogen ion activity the ratio $\frac{K}{K'}$ gives the activity coefficient of the metal ion at any particular ionic strength. The activity coefficients of Cl^- and SO_4^{--} are derived from $\frac{[Cl^-]^2}{a^2_{OH^-}} = K_1$ and $\frac{[SO_4^{--}]}{a^2_{OH^-}} = K_2$ obtained by pH measurement from reactions (1) and (2).

The occurrence of tenorite in an ore body indicates that the concentration of Cu^{++} in the solutions which are in equilibrium with the oxidized ore is less than about $10^{-4.5}$ molar.

INTRODUCTION

In the spring of 1958 we reported on a simple method of determining approximate single ion activity coefficients from pH measurements and noted the applicability of the procedure to the determination of thermodynamic properties of oxides or hydroxides (Barton and Bethke, 1958a). In the fall of the same year we discussed the application of the method to certain zinc and copper minerals (Barton and Bethke, 1958b). This report is intended to describe the method fully and to present the various data on which the above-mentioned talks were based. Specifically, data are reported on zincite for 25°C and on brochantite, atacamite, and tenorite at several temperatures between 25° and 75°C. These data are by-products of a more extensive, and as yet incomplete, study of the solubility of minerals of the heavy metals in aqueous fluids of interest in ore genesis.

GENERAL CONSIDERATIONS

The solubility of an electrolyte is expressed rigorously in terms of an activity product constant, K, in which the activities (indicated by a_i) for the

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various ions are used rather than their concentrations (indicated by parentheses).

Consider, for example, the dissolution of zincite:

$$ZnO + H_2O \rightleftarrows Zn^{++} + 2OH^- \qquad (1)$$

In infinitely dilute solutions the activities of the zinc and hydroxyl ions are equal to their concentrations and the activities of water and the solid phase are unity. Thus we may write:

$$K_{ZnO} = a_{Zn^{++}} \cdot a^2_{OH^-} = (Zn^{++}) \cdot (OH^-)^2$$

In real solutions (of finite concentration) the activity of an ion differs from its concentration by a factor dependent on various properties of the ion itself and on the total concentration of the solution. Single ion activities are not measured directly (in fact, it is thermodynamically impossible to measure the activity of a single ion in a rigorous manner). However, by making the reasonable assumptions that equilibrium is attained, and that the measured pH is a close approximation to the hydrogen ion activity, we may calculate a_{OH^-} from pH measurements and the ion product of water. The concentration of zinc may be determined analytically and hence we can define a hybrid, but measurable, constant, K′, having the form:

$$K' = (Zn^{++}) \cdot a^2_{OH^-}$$

The numerical value of K′ is a function of the composition of the solution, and approaches the thermodynamic constant, K, as the composition approaches infinite dilution; that is, $a_{Zn^{++}}$ approaches (Zn^{++}) as infinite dilution is approached. It is common practice in chemical work to represent the composition of a solution in terms of ionic strength, μ, where:

$$\mu = 1/2\Sigma\, m_a \cdot Z_a^2 + m_b \cdot Z_b^2 + \ldots m_i \cdot Z_i^2$$

m_i being the molality of the i^{th} ion and Z_i its charge. The concept of ionic strength is useful in this case because, as predicted (for dilute solutions) by the Debye-Hückel theory, log K′ plotted against $\sqrt{\mu}$ gives a straight line which may be extrapolated to infinite dilution to estimate K (for details see any physical chemistry textbook).

To show this more clearly we shall consider the titration, at equilibrium, of a Zn^{++}-bearing solution with a strong base to precipitate zincite. The resulting titration curve will be similar to one of those shown in figure 1. To the extent that the buffering action of water can be neglected, the first microdrop of base will raise the pH to the point where the solubility product of zincite is reached and precipitation begins. Further addition of base traces out the titration curve. If the same titration is rerun using the same concentration of Zn^{++} but with increased ionic strength (produced by adding NaCl to the solution, for example) precipitation will begin at a higher pH and the curve will lie to the right of the original curve, but will be identical in shape. In dilute solutions the shape and position of the entire curve are predictable on the basis of a knowledge of K′ (the change in ionic strength due to the addition of base and the removal of Zn^{++} being minor) and conversely, from any single point on the curve, K′ can be calculated if the concentration of the zinc is known. If a number of such titrations are run, a series of values of K′ can be calculated

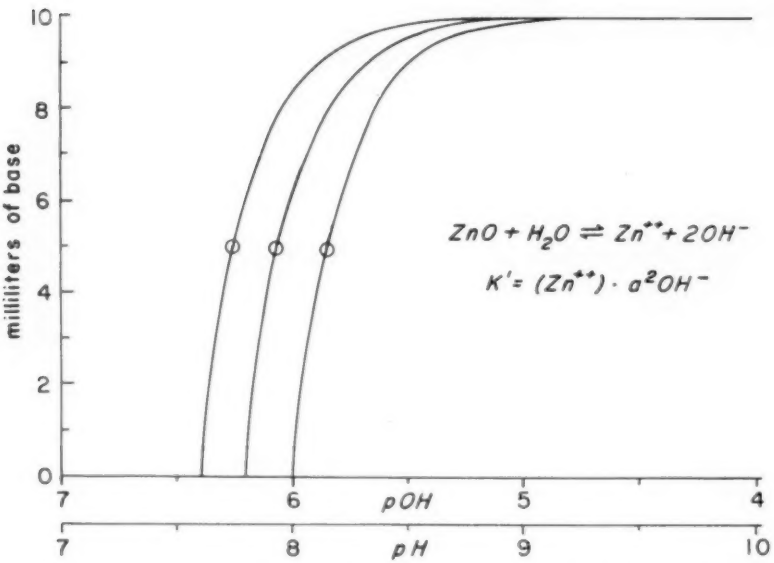

Fig. 1. Typical titration curves for the precipitation of ZnO at different ionic strengths. Circles indicate mid points of titration curves.

and plotted against $\sqrt{\mu}$. Extrapolation of these curves to infinite dilution then defines the thermodynamic equilibrium constant. However, because a knowledge of the position of any point on the curve plus the concentration of Zn^{++} is sufficient to calculate K′, it is not necessary to run the entire titration curve, and only a single point (such as those indicated by the circles in figure 1) need be determined experimentally.

ZINCITE EXPERIMENTS

Several (usually 15 to 25) beakers each containing 10 ml of 0.01 m $ZnSO_4$, 10 ml of 0.01 m NaOH, and 180 ml of NaCl solution of varying concentration were placed in a water bath. The amount of base was enough to precipitate exactly half of the zinc as zincite, and thus the final concentration of Zn^{++} in all the runs was 0.00025 m. Similar runs were made in which no NaCl was added, but instead, varying amounts of $ZnSO_4$ and NaOH were added to produce the desired changes in ionic strength. Both types of preparation gave similar results at the low ionic strengths considered in this report.

The experiments were carried out in water baths whose temperatures were controlled to within ± 2°C of those reported.

Solutions were prepared from analytical grade reagents. The water used was distilled, cooled, passed through a deionizing resin, and stored in Pyrex under a CO_2-free atmosphere. Concentrations are reported as molalities.

All runs were made in 300 ml Pyrex beakers. A thin film of mineral oil was placed on top of the solutions to prevent evaporation and the entrance of

CO_2 from the atmosphere. Although CO_2 is appreciably soluble in mineral oil, the following experiment showed that the oil definitely protects the solution from CO_2. Two aliquots of an alkaline $BaCl_2$ solution were placed in similar beakers; one was protected by oil, the other not. Both were slowly stirred. In a few hours a crust of $BaCO_3$ formed in the unprotected beaker, while no $BaCO_3$ was visible even after a week in the other. The buffer solutions were protected by oil without measurable effect on their pH. A film of oil on the surface of the bath also made it possible to use water baths up to at least 95°C without excessive evaporation.

The pH was measured using a Beckman general purpose glass electrode against a calomel electrode with a saturated KCl bridge having a permeable-fiber type junction with the solution. In most measurements no sodium ion correction was necessary. The electrodes were rinsed with acetone (to remove oil) and then with water between measurements. Wetting the electrodes with water and then quickly plunging them through the oil layer into the solution usually prevented the oil from fouling the electrodes.

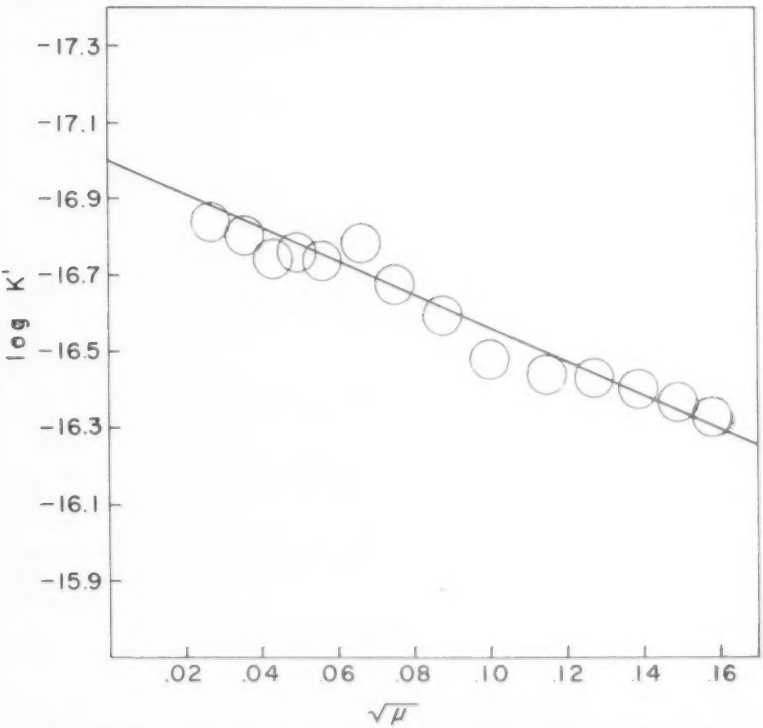

Fig. 2. Extrapolation of $K' = (Zn^{++}) \cdot a^2_{OH^-}$ to infinite dilution. Size of circles indicates magnitude of the uncertainty of individual pH measurements.

A Beckman model H-2 meter was used, and standardization was against three Beckman buffers having pH values (at 25°C) of 4.00, 7.00, and 10.00 with an uncertainty of 0.01. The buffers were placed in the bath with the runs prior to standardization and the pH of the buffer at the temperature of the run was read from a graph plotted from data supplied with the buffer. It was usually possible to check all three buffers to within ± 0.02 pH units. If the meter was not turned off, the drift usually was less than 0.05 pH units in 24 hours.

The pH measurements were made at intervals of from a day to a week. The consistency of the pH values obtained after a week or two is the principal criterion for the attainment of equilibrium. From the pH values and the known concentration of Zn^{++}, K′ was calculated for each ionic strength. A plot of log K′ against $\sqrt{\mu}$ shown in figure 2 extrapolates as a straight line to log K = 17.0. For zincite and for the later described copper compounds, the curves were extrapolated to infinite dilution strictly on the basis of the experimental points at the temperature considered, without regard for the slopes of similar curves at other temperatures or for the theoretically derived Debye-Hückel limiting slope. The result is to give a little more scatter to the log K against 1/T plot than might be otherwise obtained. The most dilute runs in several of the copper as well as zinc experiments gave anomolously low K′ values which are presumably due to minor amounts of CO_2, hence the curves were weighted in favor of the points at higher ionic strengths.

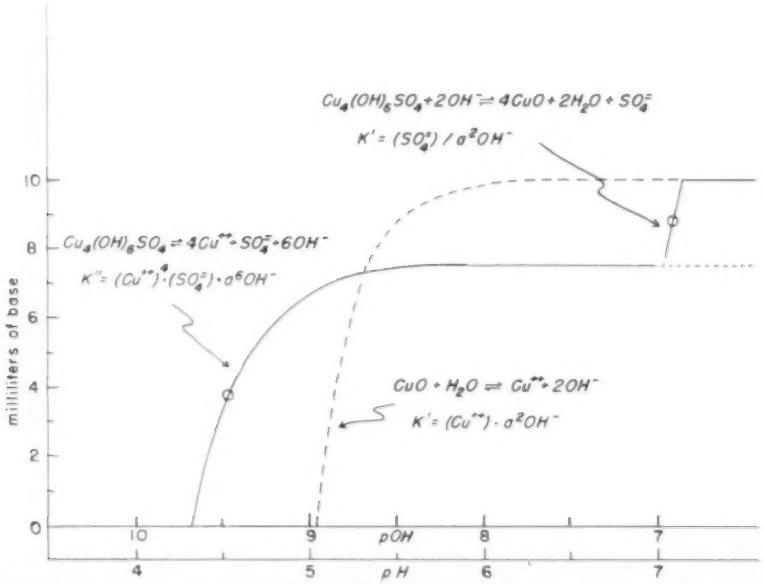

Fig. 3. Typical titration curve for cupric sulfate solutions. Equilibrium curves solid; metastable curves dashed. Circles indicate mid points for each reaction.

On completion of the runs the precipitates were spot checked by x-ray diffraction to identify the phases. Most frequently the solid phase was zincite, but there were some randomly scattered occurrences of zinc hydroxides.

ATACAMITE AND BROCHANTITE EXPERIMENTS

The titration of cupric solutions with a strong base is more complicated because of the formation of basic salts. In dilute cupric chloride and cupric sulfate solutions one stable basic salt was observed in each, respectively atacamite, $Cu_4(OH)_6Cl_2$, and brochantite, $Cu_4(OH)_6SO_4$. Very dilute or very basic solutions gave tenorite, CuO. The equilibrium titration curves and the reactions taking place along them are shown in figure 3. There are two precipitation curves and a reaction segment between them. The composite stable curve is shown as a solid line; the metastable portions are dashed. The first addition of base to $CuSO_4$ solution initiates the precipitation of brochantite according to the reaction:

$$Cu_4(OH)_6SO_4 \rightleftarrows 4Cu^{++} + 6\,OH^- + SO_4^{--} \qquad (2)$$

for which,

$$K = a^4_{Cu^{++}} \cdot a^6_{OH^-} \cdot a_{SO_4^{--}}$$

Bronchantite continues to precipitate with addition of base until essentially all the copper has precipitated. Further addition of base rapidly increases the pH until a critical $\frac{a_{SO_4^{--}}}{a^2_{OH^-}}$ ratio is reached at which point the ratio remains constant while brochantite reacts with OH^- to yield tenorite and SO_4^{--} by the reaction:

$$Cu_4(OH)_6SO_4 + 2\,OH^- \rightleftarrows 4CuO + SO_4^{--} + 4H_2O \qquad (3)$$

for which,

$$K = \frac{a_{SO_4^{--}}}{a^2_{OH^-}} .$$

The slight slope to this reaction segment is due to the release of SO_4^{--} during the reaction, thereby requiring slightly more OH^- to maintain the proper $\frac{a_{SO_4^{--}}}{a^2_{OH^-}}$ ratio. If more dilute solutions are used both precipitation curves are shifted to the right (toward higher pH), but the brochantite curve moves more rapidly than the CuO curve and thus eventually the solution will become sufficiently dilute for tenorite rather than brochantite to be the initial phase precipitating.

Because the equilibrium constant is predictable on the basis of any known point on the curve, only one point for each reaction was run. The positions of the experimental points from which the equilibrium constants were determined

are shown by the circles in figure 3. The runs were made up by adding just enough NaOH to a series of solutions having variable amounts of $CuSO_4$ to carry the titration to the mid points of the two curves (this amounted to 37.5 and 87.5 percent of the amount of base necessary to completely precipitate the copper as CuO). The details of the experimental procedure are similar to those described for zincite.

Figure 4 shows the extrapolation of the constant for brochantite, $K''_{brochantite} = (Cu^{++})^4 \cdot (SO_4^{--}) \cdot a^6_{OH^-}$,[1] to infinite dilution to determine

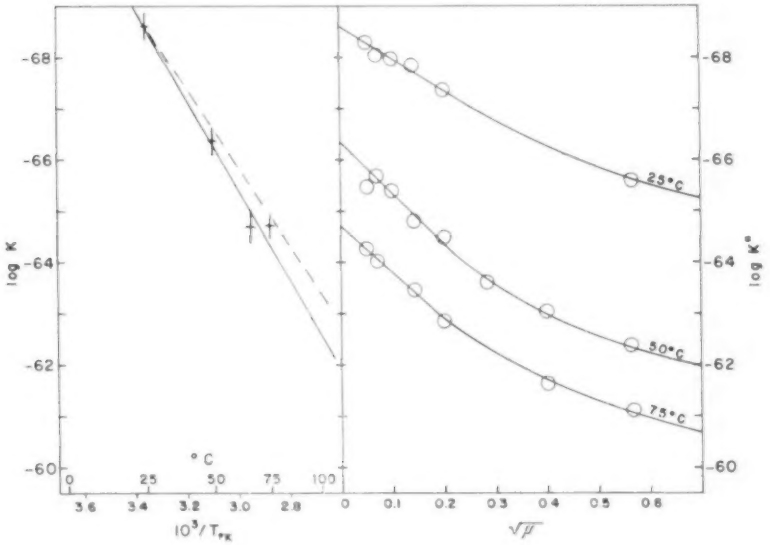

Fig. 4. Extrapolation of $K'' = (Cu^{++})^4 \cdot (SO_4^{--}) \cdot a^6_{OH^-}$ to infinite dilution for various temperatures (right), and plot of $K = a^4_{Cu^{++}} \cdot a_{SO_4^{--}} = \cdot a^6_{OH^-}$ against reciprocal absolute temperature (left). Size of circles indicate magnitude of uncertainty of individual pH measurements. + indicate magnitude of estimated ± in log K.

the thermodynamic constant, K, at several temperatures. Also shown is the plot of log K against reciprocal absolute temperature. The dissociation constants for water used in determining a_{OH^-} from the pH at various temperatures are taken from Bjerrum (1929).

The experiments with atacamite were similar in every respect to those with bronchantite except that $CuCl_2$ rather than $CuSO_4$ was used in the initial

[1] K'' is used to note a hybrid constant involving two concentration terms in contrast to K' which is used to denote hydrid constants involving only one concentration term.

solutions. The titration curves have exactly the same shape but different positions and are based on the reactions:

$$Cu_4(OH)_6Cl_2 \rightleftarrows 4Cu^{++} + 6\ OH^- + 2Cl^- \qquad (4)$$

for which

$$K = a^4_{Cu^{++}} \cdot a^6_{OH^-} \cdot a^2_{Cl^-}$$

and

$$Cu_4(OH)_6Cl_2 + 2\ OH^- \rightleftarrows 4CuO + 2Cl^- + 4H_2O \qquad (5)$$

for which

$$K = \frac{a^2_{Cl^-}}{a^2_{OH^-}} .$$

The extrapolation of $K''_{atacamite}$ to infinite dilution to define the thermodynamic constant, K, at several temperatures is illustrated in figure 5 along with a plot of log K against reciprocal absolute temperature.

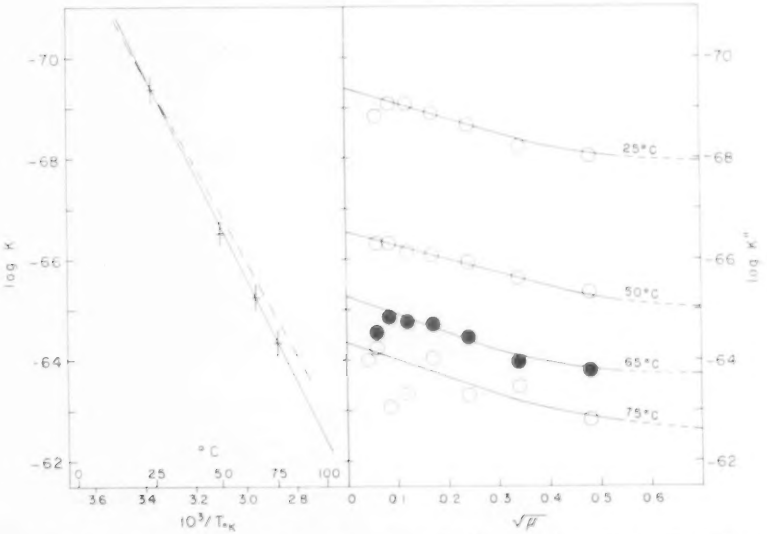

Fig. 5. Extrapolation of $K'' = (Cu^{++})^4 \cdot (Cl^-)^2 \cdot a^6_{OH^-}$ to infinite dilution for various temperatures (right), and plot of $K = a^4_{Cu^{++}} \cdot a^2_{Cl^-} \cdot a^6_{OH^-}$ against reciprocal absolute temperature (left). Size of circles indicates magnitude of uncertainty in individual pH measurements. + indicates estimated ±in log K.

By subtracting reaction 3 from reaction 2, or reaction 5 from reaction 4, the reaction for the dissolution of tenorite is obtained:

$$4CuO + 4\ H_2O \rightleftarrows 4Cu^{++} + 8\ OH^- ,$$

or

$$CuO + H_2O \rightleftarrows Cu^{++} + 2\ OH^- ; \qquad (6)$$

for which:

$$K = a_{Cu^{++}} \cdot a^2_{OH^-} .$$

In figure 6, K′ for reactions 3 and 5 are plotted against $\sqrt{\mu}$ and the thermodynamic constants, K, are determined by extrapolation to infinite dilution. The numerical values of $K_{tenorite}$ are obtained at various temperatures by dividing the above values of K into those for reactions 2 and 4. These values of $K_{tenorite}$ are plotted against reciprocal absolute temperature in figure 7.

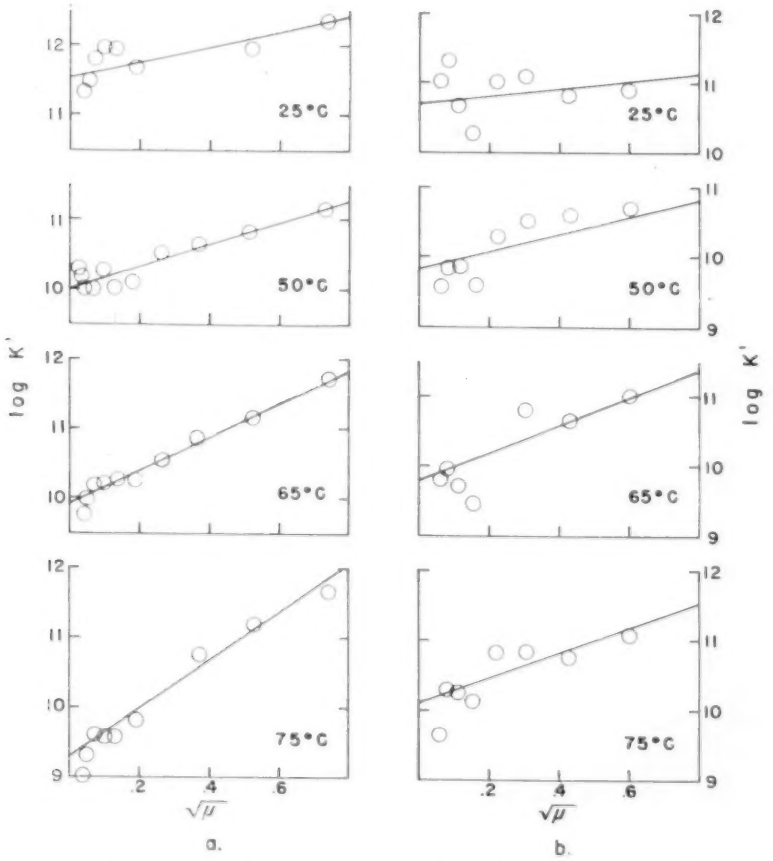

Fig. 6. a. Extrapolation of $K' = (SO_4^{--})/a^2_{OH^-}$ to infinite dilution at various temperatures, b. Similar extrapolation of $K' = (Cl^-)^2/a^2_{OH^-}$. Size of circles indicates magnitude of uncertainty in individual pH measurements.

The precipitates from both the bronchantite and atacamite runs were examined by x-ray diffraction techniques and were always brochantite, atacamite, or tenorite as predicted. The length of time necessary for the attainment of equilibrium ranged from a week or two at 25°C to a few hours at 75°C.

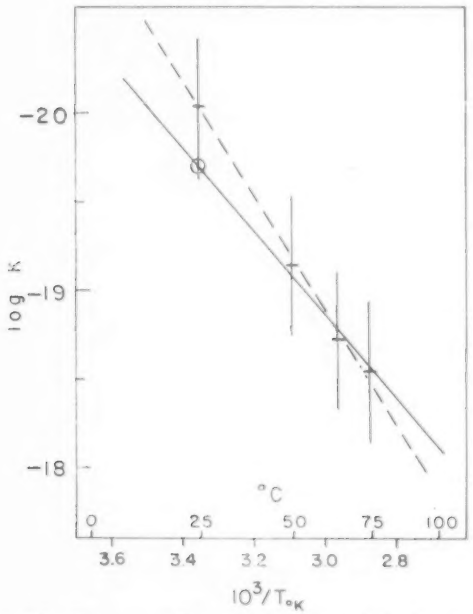

Fig. 7. Plot of log K = $a_{Cu^{++}} \cdot a^2_{OH^-}$ against reciprocal absolute temperature. + indicates estimated ± in log K. Circled point at 25° is that of Näsänen and Tamminen.

THERMODYNAMIC CALCULATIONS AND COMPARISON WITH THE RESULTS OF OTHERS

The free energy changes for the various reactions may be calculated from the relation:

$$\Delta F = -RT\ln K$$

By combining the free energy changes so calculated with the standard free energies (from Latimer, 1952) of the remainder of the products and reactants we calculate the standard free energies for the synthetic minerals. The data are summarized in table 1.

The enthalpy changes for the reactions can be calculated in two ways. The first is to apply the relation:

$$\frac{d\ln K}{d(1/T_K)} = \frac{\Delta H}{R} \qquad (a)$$

to the log K versus 1/T plots in figures 4, 5, and 7. The second is to calculate the entropy changes for the various reactions and then to combine these with the free energies according to the equation:

$$\Delta F = \Delta H - T\Delta S \qquad (b)$$

The heats of reaction are then combined with the standard heats of formation of products and reactants to calculate the unknown standard heats of formation. The results of both methods of calculation are given in table 1.

TABLE 1

Thermodynamic data for some synthetic zinc and copper minerals at 25°C

(Better values are underlined. Estimated entropies are in parentheses.)

Mineral	K_{298}	$\Delta F°_{298}$ (Kcal/mole)	$\Delta H°_{298}$ (Kcal/mole)		$S°_{298}$ (cal/mole degree)	
			Calculated from $\frac{d\ln K}{d(1/T)} = \frac{\Delta H}{R}$	Calculated from $\Delta F°$ and the estimated $S°$	Calculated from $\Delta F°$ and the first $\Delta H°$	Estimated or measured by others
Zincite	$10^{-17.0 \pm 0.1}$	-76.9 ± 0.15	—	**-84.0**	—	**10.5**
Brochantite	$10^{-69.6 \pm 0.4}$	-434.3 ± 0.6	**-525.4**	-520.9	**72.3**	(87.4)
Atacamite	$10^{-69.4 \pm 0.4}$	-320.7 ± 0.6	**-396.2**	-392.1	**75.2**	(86.4)
Tenorite	$10^{-10.7 \pm 0.4}$	-29.8 ± 0.6	**-40.4**	**-36.9**	**-1.3**	**10.19**

The ± attached to the ΔF° values in table 1 reflects only the ± in our measurement of K and does not include any allowance for uncertainty in Latimer's data.

The value of log $K_{ZnO} = -17.0 \pm 0.1$ is in excellent agreement with the results of Feitknecht (1947) who gave –17.036 and –17.075 respectively, as the logarithms of the solubility product constants for zincite and ε $Zn(OH)_2$, the most stable crystalline forms at 25°C. The ΔH° calculated from relation (b) using Latimer's (1952) value for the entropy of ZnO is –84.0 Kcal and is in agreement with the –83.25 Kcal reported by Coughlin (1954).

Näsänen and Tamminen (1949) determined K″ for brochantite and atacamite at 25°C and their data extrapolated to infinite dilution yield values of log K = –68.40 and –69.00 for brochantite and atacamite respectively. Silman[2] has measured the log K for brochantite as –68.8. All these values were determined by experiments similar to ours and are in excellent agreement with our findings. To our knowledge the heats of formation and entropies of brochantite and atacamite have not been measured previously. The entropies of brochantite and atacamite estimated by summing the entropy contributions of the constituent ions according to the method of Latimer (1952) are 87.4 and 86.4 respectively, in fair agreement with those calculated from our experimental data. Because of the approximate nature of the estimated entropies, more reliability is placed on the values of ΔH° calculated from our measurements than those calculated from the estimated entropies. The curves given by our experimental data are shown as solid lines in figures 4 and 5; the curves derived from the estimated entropy values are dashed in for comparison.

Näsänen and Tamminen (1949) measured the solubility of tenorite directly in dilute perchlorate solutions and give a K corresponding to a ΔF° of –29.9 Kcal. Latimer (1952) and Coughlin (1954) give $\Delta F^\circ = -30.4$ Kcal and –31.0 ± 2.0 Kcal respectively. Recalling that the values for tenorite in table 1 were derived from two pairs of experimentally determined constants (the combination of the K for atacamite with the K for the atacamite-tenorite reaction and the K for brochantite with the K for the brochantite-tenorite reaction) the agreement between table 1 and the literature is very gratifying. Although the ΔH° of tenorite calculated from our data (dashed line in fig. 7) is –40.4 Kcal in fair agreement with the –37.1 Kcal of Latimer and –37.6 ± 2.0 Kcal of Coughlin, the calculated entropy of –1.3 is in very poor agreement with the measurement of 10.19 by Hu and Johnston (1953). However, if the K at 25° of Näsänen and Tamminen (which lies within the uncertainty of our determination) is used instead of our value, the results may be reconciled. The solid line in figure 7 which has a slope determined by Hu and Johnston's entropy value and Näsänen and Tamminen's K fits our points fairly well, hence we consider Näsänen and Tamminen's result to be the better value.

PRECIPITATION OF TENORITE IN NATURAL ENVIRONMENTS

Copper sulfate is probably the most likely soluble form of copper to expect

[2] Ph.D. thesis, Harvard University, 1958.

in an oxidizing sulfide ore body. From the K of tenorite and of brochantite we can easily calculate that if the concentration of $CuSO_4$ were greater than about $10^{-4.5}$ moles per liter (or about 2 parts per million Cu^{++}) brochantite will be stable rather than tenorite. Higher concentration of sulfate will only increase the relative stability of brochantite; and if other geologically reasonable acid-forming anions, such as chloride, phosphate, carbonate, and so forth, are used either in place of sulfate in dilute solution, or as salts (that is, NaCl or $CaCl_2$) in higher concentration to raise the ionic strength and thereby increase the concentration of Cu^{++} by lowering the activity coefficient, they too will form stable, insoluble compounds with Cu^{++}. Hence, the occurrence of tenorite indicates that the concentration of Cu^{++} in the solutions was very low. Garrels and Dreyer (1952) have used similar reasoning to conclude that malachite also indicates a very low concentration of Cu^{++}.

SINGLE ION ACTIVITY COEFFICIENTS

In an earlier abstract (Barton and Bethke, 1958a) we noted a method for determining approximate single ion activity coefficients from pH measurements. Using zincite as an example, we note that $K = a_{Zn^{++}} \cdot a^2_{OH^-}$ and that $K' = (Zn^{++}) \cdot a^2_{OH^-}$. On dividing the first equation by the second the $a^2_{OH^-}$ cancels out and hence $\frac{K}{K'} = \frac{a_{Zn^{++}}}{(Zn^{++})}$ which in turn equals the single ion activity coefficient for Zn^{++}. In a like manner the atacamite-tenorite reaction can give the activity coefficient of Cl^- and the brochantite-tenorite reaction can give the activity coefficient for SO_4^{--}. The activity coefficient for Cu^{++} can be evaluated from $\frac{K}{K''}$ for atacamite (provided that the activity coefficient for Cl^- is also measured in a solution of essentially the same composition). This method offers a simple, broadly applicable way to study the solubility of compounds and thereby to study complex formation. In our work the experimental uncertainty in measuring pH makes the determination of activity coefficients less precise by perhaps two orders of magnitude than conventional methods for the determination of mean activity coefficients in simple solutions, however, the method described here does have the advantage of being able to provide data from very complex solutions. The major requirements are that an oxide, hydroxide or basic salt be stable in the solutions and that reasonably reliable pH measurements can be made.

Lefebvre (1957) has devised a method in which the areas below (or, as plotted in our diagrams, to the left of) a series of titration curves are measured for different concentrations of reacting components but with ionic strength maintained nearly constant by some "inert" electrolyte. He has applied this method to the study of polymerization and complex formation, but has not tried to estimate single ion activity coefficients.

ACKNOWLEDGMENTS

Most of the actual measurements whose results are described in this report were made by David Daniels, Norman Page, Chester Nichols, and Miss Holly Edwards.

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GAS AS A SEDIMENTARY AND DIAGENETIC AGENT*

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ABSTRACT. Gas can be a significant factor in the transportation and alteration of sediments, and traces of its former presence may convey useful information about the depositional and diagenetic history of sedimentary rocks. Transportation of sediment by gas is probably most important in embayed littoral environments of broad extent and reduced offshore drainage, hence onshore transport and shoreward decreasing grain size. Stranded, blown, or drifting bubbles and evasion of upwelling gas are responsible for a variety of imprints and structures in sediments. The relations to other features of impressions of gas blisters at the base of a thin Devonian graywacke are analyzed and found to support deposition of the graywacke from a weak suspension current. Concentration of pyrite in irregularly subcylindrical structures within a Devonian limestone contributes to the interpretation of these structures as former sites of gas accretion. Relations between pyrite, matrix, and later mineralization reveal the main steps in the diagenetic history of the limestone. Some ancient sedimentary features that have been interpreted as raindrop imprints, tracks and burrows, and Precambrian jellyfish are probably also the work of gas.

INTRODUCTION

The importance of gas in transporting, marking, and altering sediments, and of its traces as clues to depositional, paleoecological, and diagenetic history is not generally appreciated. Despite the frequency with which stagnant basins and anaerobic conditions are invoked in geologic reconstruction, and although it seems intuitively probable that the effects of gas on sedimentary rocks should be both widespread and recognizable, little has been published that deals with specific gas effects in the stratigraphic column.

Both above and below existing depositional or sediment-water interfaces, however, dissolved and even free gas is common. Interstitial gases are organic and primarily bacterial in origin. They consist mainly of carbon dioxide, methane, ammonia, hydrogen sulfide, and hydrogen. Gas in the water above consists of unbalanced free biologic oxygen within the photic zone, dissolved carbon dioxide, and such bacterially generated gases of other sorts as may rise above the depositional interface, result from interactions enroute, or issue from decompositional or digestive processes within the water mass. Such gases may be abundant locally, and biologic or atmospheric gases are generally entrained in forthy mixtures in intertidal waters. These gases alter the physical and chemical properties of their surroundings and aid in the transport of particulate materials.

The abundance of gas in natural sediments and waters under suitable circumstances is easily ascertained by observing its evolution in clear containers of fine-grained sediment, the "boiling" of certain ponded waters when hydrostatic head above their sediment floor is lowered, the frothy scum of flooding tidal flats and inlets, and the record of aquatic mass mortalities resulting from oxygen depletion and the rise of poisonous gases from the bottom sediments (Brongersma-Sanders, 1957, p. 959, and references there cited). A variety of gases is reported from the bottom sediments of deep marine basins off the coast of southern California by Emery and Hoggan (1958). Ohle (1959) describes and illustrates with striking echogram records the cyclical evasion of methane and other gases from bottom muds in North German lakes in such quantities

* Publication authorized by the Director, U. S. Geological Survey.

and with such force as to generate upwelling currents capable of transporting sediment particles across the thermocline and into the surface waters from depths of more than 30 meters. He also discusses (p. 139-146) the various interactions by which methane may be oxidized to carbon dioxide, carbon dioxide hydrogenated to methane, and methane utilized as an energy source in sulfate reduction with hydrogen sulfide as a product.

The present paper, however, is not concerned with such complexities. Its aim is simply to stress the sedimentary and diagenetic importance of gas by reviewing some aspects of the general field, and by discussing the implications of selected structures believed to have been created or modified by the action of gas in ancient sediments.

TRANSPORT OF SEDIMENT

The potential significance of sediment transport by gas is illustrated by its use in flotational ore beneficiation and by the recycling problem it creates in sewage disposal (Menard, 1950, p. 99, 105). Gas transport of sedimentary particles in nature is described by Nordenskiold, by Menard, by Bradley, by Bourcart and Roa-Morales, and by Ohle.

Nordenskiold (1900) describes the floating of quantities of shale fragments weighing up to 0.8 gram apiece as a partial result of "small gaseous bubbles attached to the under surface" in the sea off southwestern Patagonia. Menard (1950) cites experimental work by himself and observations by others to show that attachment of bubbles to grains of sand and silt provides enough buoyancy to produce or augment surface flotation, and to enhance saltational movement along the bottom. Bradley (1957, p. 665) records the shoreward transport of silt and organic matter on a Maine tidal flat in scum films produced by rising gases of biologic origin. Bourcart and Roa-Morales (1957, p. 547) conclude from detailed quantitative data (oral communication, 1956) that onshore transport in frothy surface scum is currently an important sedimentary mechanism in the Bay of Mont St. Michel, the Rance embayment, and other tidal inlets around Brittany—especially with reference to the origin of the calcareous mud called "tangue" (Bourcart et al., 1944). And Ohle (1959, p. 132-133) describes the ascension of lake-bottom sediments in gas-generated upwelling currents.

Gases of biologic and atmospheric origin thus can play a significant sedimentary role, either in silled basins or, and especially, in embayed intertidal environments of wide extent, restricted terrestrial runoff, and generally large tidal range. Such intertidal environments, owing to the influence of current variations and unusual transport factors, are also likely to be characterized by onshore movement of sediment and shoreward decreasing grain size (for references see Cloud, 1959, p. 937). In such areas too, light particles such as the tests of microorganisms, sponge spicules, and fine-grained or highly porous sediment can be floated to the surface from deeper offshore water by rising gas, and thence conveyed to tidal flat deposits in the vesicular surface scum, defying the general rule that redeposition is a purely downslope process.

SURFACE MARKINGS

Gas-produced surface markings are of two principal classes: *evasion marks*, due to localized effusion or temporary confinement at the sediment surface of gas generated at a lower level, and *contact marks*, attributable to bubbles stranded on, drifted across, or blown along the sediment surface. Gas evasion marks and gas contact marks each include a variety of distinctive subordinate markings which, although now rarely reported from the geologic record, were in some degree familiar to our classical predecessors (e.g. Buckland, 1842; Lyell, 1851, p. 241; James, 1884, p. 128-129; Burroughs, 1924).

The best places to observe the formation and range of markings actually being produced by gas evasion are in the mud pots of thermally active areas and in areas of cyclically inundated organic-rich muds or muddy silts. Discounting wide differences in size and rate of development, the process is much the same at either place.

First a simple mound or blister develops (Goeman, 1939; Schäfer, 1939, fig. 4; Reineck, 1955, pl. 8, fig. 5). Collapse then commonly occurs, leaving a doughnut-shaped rimmed depression or dimpled blister (Klippel, 1939, especially figs. 1-3). Next a second generation blister may rise within this rimmed depression to produce a concentric double blister, or fried-egg pattern (pl. 1A). In mud pots, successive blisters within blisters commonly form multiple concentric rings (pl. 1A), but ordinary gas effusion from sediments normally produces few such rings about any given center. If subsidence and escape of gas continues, particularly if mixed with or supplemented by water-expulsion, or if the sediment is sandy, a small or large conical depression may form, with or without a central orifice or multiple vents (Maxson, 1940). Larger gas mounds may fracture tensionally in a radial or irregular pattern, or may give rise to gas "calderas" by complete disruption or subsidence (Schäfer, 1954). Where the sediment has a high degree of consistency, some of the initial gas blisters may blow out or be pierced by an apical pore as a result of forceful gas evasion (Reineck, 1955, pl. 8, fig. 5; Twenhofel, 1921, fig. 7). They then resemble the genetically different pit and mound structures of Kindle (1916), Twenhofel (1932, p. 680-682), and Shrock (1948, p. 132-135), which result from convectional discharge of water from settling clayey sediments (Kindle, 1916; Schofield and Keen, 1929), and which form surface elevations when the initial dilution is great and surface depressions (or "pit in pit" structure) in more thixotropic mixtures (Schofield and Keen, 1929). Pierced gas blisters can be distinguished from pit and mound structure, however, by the greater irregularity of their central pores and by the fact that they invariably are associated with proportionally large numbers of simple blisters, whereas pit and mound structures should all have central pores (illustrations in references cited). Similar structures may result from the forceful expression of entrapped air in beach sands of suitable composition (*e. g.* Emery, 1960, p. 193-195).

Most, but not all, of the features described are matched by known occurrences in the geologic record. Surface markings that strikingly resemble the simple initial gas blister, the succeeding rimmed depression and dimpled

blister, and the concentrically blistered stage are found in the Cambrian of Arizona (McKee, 1945, p. 131, pl. 7d) and the Devonian of Pennsylvania (pl. 1B-E). Pierced blisters are not known, although some of the dimpled blisters illustrated in plate 1D might have had central pores at one time. Structures on the pitted surface of a Carboniferous slate from Massachusetts, considered by Clark (1923, p. 482-484, fig. 2) to be the work of gas, are probably correctly interpreted by Shrock (1948, p. 134-135, fig. 95) as true pit and mound structure, and Madigan (1928) describes probable pit in pit structure from the Point Marsden series (Cambrian) of Australia. Deeply conical fossil gas, or gas-water, craters such as illustrated by Maxson (1940) from fresh-water silty mud flats have not to my knowledge been recorded from ancient sediments, although they surely exist. A possible ancient parallel to ruptured gas mounds will be discussed on a following page.

Stranding, drifting, or blowing of gas bubbles at the sediment surface produces distinctive contact marks. Where gas works upward from below to form bubbles on the surface of the sediment, or where bubbles are generated at the sediment surface or carried down to it from above and temporarily anchored there by attached particles, the surface tensional effects or bursting of such bubbles may create imprints resembling those attributable to rain drops (Buckland, 1842; Desor, 1850; Lyell, 1851, p. 241-243; Hughes, 1884, p. 183, pl. 11; Martin, 1904; Twenhofel, 1921, p. 365-366, fig. 5; Twenhofel, 1932, p. 682-683; Shrock, 1948, p. 143, 145; Reineck, 1954, 1955). The differences between bubble, raindrop, and hail imprints were carefully analyzed by Lyell and later by Twenhofel in the works cited. Imprints of stationary bubbles differed from raindrop imprints in being highly symmetrical, in lacking raised rims, and in having depth/width ratios greater than those of rain imprints. Ideally their profiles are arcs of circles. Martin notes that, under certain conditions, bubbles in contact with sediment but covered by only a thin film of water and blown by the wind may produce marks that are ovate in plan, develop well-defined margins, and, of course, become larger and tend to assume a preferred orienation. Gas bubbles or foam masses may also drift or blow along the sediment surface so as to produce curving (Twenhofel, 1921, p. 367, fig. 6) or straight tracks or distinctive patterns of parallel lines. Simple and compound rectilinear patterns produced by wind-blown bubbles have not only been produced experimentally, they are also duplicated in the upper Permian Kreuznacher Schichten of Germany (Reineck, 1954, especially figs. 1-6).

Surface markings produced by gas can, thus, tell us interesting and perhaps useful things about sediment chemistry and microbiology, depth of water, and even the meteorology of ancient environments. A detailed awareness of their nature can prevent the paleoecological misconstructions that would result from confusing them with unrelated features.

The gas blisters illustrated on plate 1B-E provide an unforseen sedimentological application. These features are preserved in reverse orientation on the bottom surface of a four-inch bed of fine-grained graywacke in an Upper Devonian graywacke-shale sequence in southeastern Pennsylvania. This se-

quence was shown to me by Professor Francis Pettijohn and his students, Norman Kiever and Earl McBride, who interpret it (oral communication, 1959) as a flysch-facies deposit in which the graywacke beds were emplaced by turbidity currents. The gas markings presumably could not have developed in the ideal form they possess except under uniform surface tension, therefore not at a sand-shale boundary within the sediment but only at a sediment-water (or wet sediment to air) interface. The delicately preserved imprints of these supposedly antecedent gas markings on the under surface of the sandstone (graywacke) bed seem to eliminate the possibility of its deposition by normal tractional processes and imply settling from suspension. Locally on this same bottom surface, however, are groups of fine parallel lines which are probably current markings, and in the same orientation as these markings is a delicate current bedding which shows that the suspended sediment had perceptible forward motion and was therefore current-borne very near or at the bottom. Where the fine parallel lines occur, they blur or obliterate the gas blisters. The fact, however, that at most places abrasion was too weak to destroy or significantly to deform the gas markings, combines with the almost uniformly fine grain-size of the sediment (exclusive of occasional crinoid columnals in stable orientation in the lower part) to suggest that the site of deposition was far enough out in a basin for the current to have been spent and the coarser fraction of its original load already dropped. On the other hand, the fact that gas was present in a free and undissolved state in bubbles up to several millimeters across may be interpreted to suggest (depending on the kind of gas) that depth, although surely well below wave base, was not profound.

INTERNAL STRUCTURES AND DIAGENETIC EFFECTS

Internal pockets and passageways made by gathering and migrating gas have previously been described from modern but not from ancient deposits. Twenhofel (1921, p. 368-369) found experimentally that gas rising through mud could produce and maintain open tubes which "resemble those made by worms" (*see also* Reineck, 1955, pl. 8, fig. 5), and Maxson (1940) observed the openings of presumably down-reaching gas tubes in river and lake sediments.

Figure 1 illustrates a pattern of gathering and moving gas formed along the liner-sediment contact in a sealed core of argonite mud from the Bahama Banks during the first month after collecting. Were similar gas-filled openings to form within a sediment column and to persist during compaction, later filling with clear calcite could produce a pattern resembling that of certain aphanitic limestones with ramifying calcite stringers, pockets, or vugs. A similar pattern, to be sure, could be produced by the ramifying angular fissures that accompany expulsion of water from settling clay-sized sediments (Schofield and Keen, 1929).

For an example of probable gas trackways in the geologic record let us consider a Devonian (Givetian) limestone from the Meuse Valley in Belgium, where, guided by Prof. Marius Lecompte, I was shown the bed from which came the specimen illustrated on plates 2 and 3. Although, in the field, the

ramifying subcylindrical structures in the weathered vertical section of this bed (pl. 2A) seemed clearly to represent a primary feature of the rock, their marked irregularity caused misgiving about categorically interpreting them as simple burrows. Polished surfaces showed that most of these markings were strikingly irregular to even ragged in longitudinal profile (pl. 2B, pl. 3A-B) but rather more regular in transverse profile (pl. 3C). Inspection of these polished surfaces (*e.g.* pl. 3B) indicated irregular marginal gradation from marking to matrix, but showed sharp distinction in the generally higher concentration of pyrite (shiny metallic grains in pl. 3A-B) in and lighter color of the markings. Thin sections (pl. 3D-E) reveal that the lighter color and distinctive weathering of the markings is due to the fact that their bulk grain composition is a fine-grained sparry dolomite. In a section sufficiently thin to illuminate the very dark matrix (pl. 3E), it can be seen that this slightly dirty sparry dolomite fades out into the surrounding rock. It is also intergrown with and includes unaltered islands of dark calcitic material identical to the surrounding matrix (pl. 3A, just left of center; also parts of 2B). It is sharply different from, although its dolomitization may be contemporaneous with, a vug-filling, white, coarse-grained, dolomite and calcite that is younger than, and not a replacement of the surrounding matrix. Cross cutting both of these, and younger than either, is a white, fibro-granular, vein calcite.[1] In addition to the pyrite, there is locally some limonite-after-pyrite within the markings.

Most of the markings are elongated in a vertical direction, but their long

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PLATE 1

Gas evasion marks on Devonian and modern sediments

A. Compound gas blisters in mudpot adjacent to geyser field at Rotorua, New Zealand (photograph by H. S. Ladd, 1926). Individual blisters about 8 inches in diameter.

B-C(× 1.4) and D-E(× 1.8). Rubber facsimiles of gas blisters from inverted impressions on under surface of an Upper Devonian graywacke, Trimmers Rock sandstone of Willard, 13 mi. NNW of Lebanon, Pennsylvania, and midway between Inwood and Twin Grove Park, in railway cut east of State highway 72 (field loc. 2 of 9 April 1959).

PLATE 2

Probable former gas pockets in a Middle Devonian black limestone

A(× 0.7). Vertical weathered section in Givetian limestone from ridge within old fort on east bank of Meuse River, opposite Givet, Belgium (field loc. 3a of 6 June, 1957).

B(× 0.7). Vertical polished section of same block as A (large pale areas at top right and top center are reentrants of weathered rock surface).

PLATE 3

Detail of same probable gas accretion sites as illustrated in plate 2

A(× 2.7). Vertical polished section counterpart to area just right of center of plate 2B.

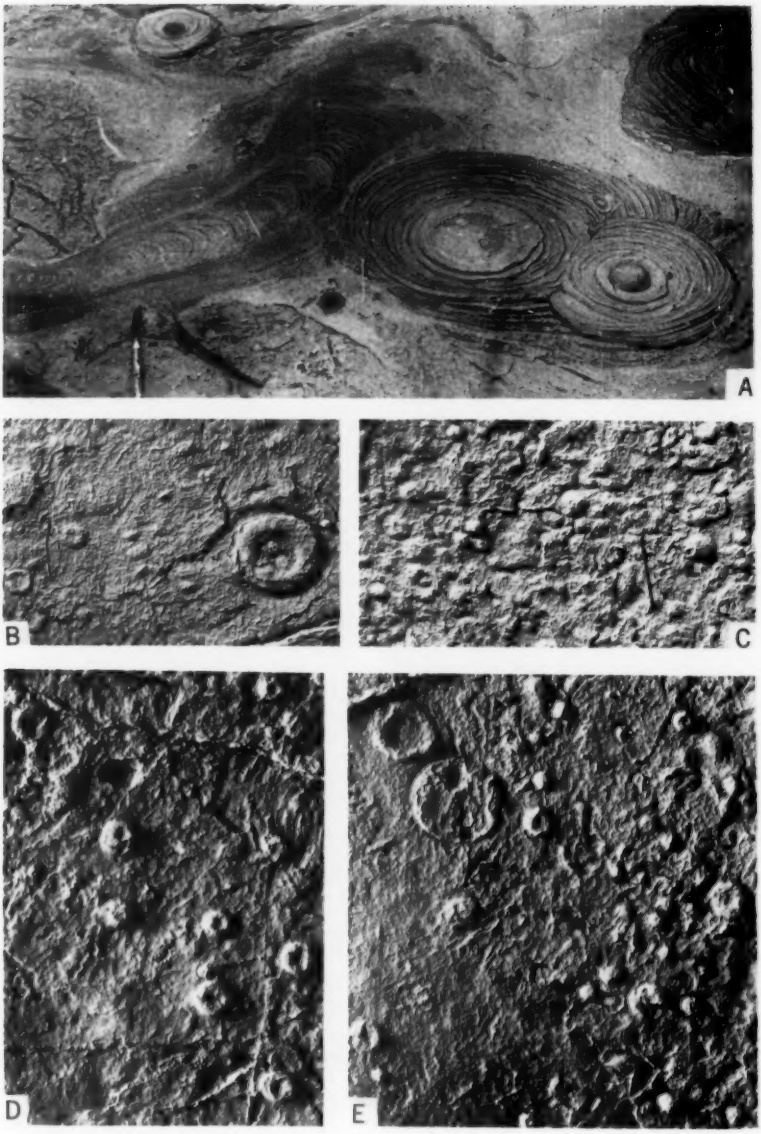
B(× 2.7). Part of same polished surface as illustrated on plate 2B, showing marked pyrite concentration in irregularly subcylindrical structure and ragged, gradational contact between this and surrounding matrix.

C(× 0.7). Horizontal polished section at base of block showing plan outline of structures.

D(× 4). Vertical thin section (crossed nicols) showing relations between sparry dolomite of pockets P and vugs V, and calcite of joints J.

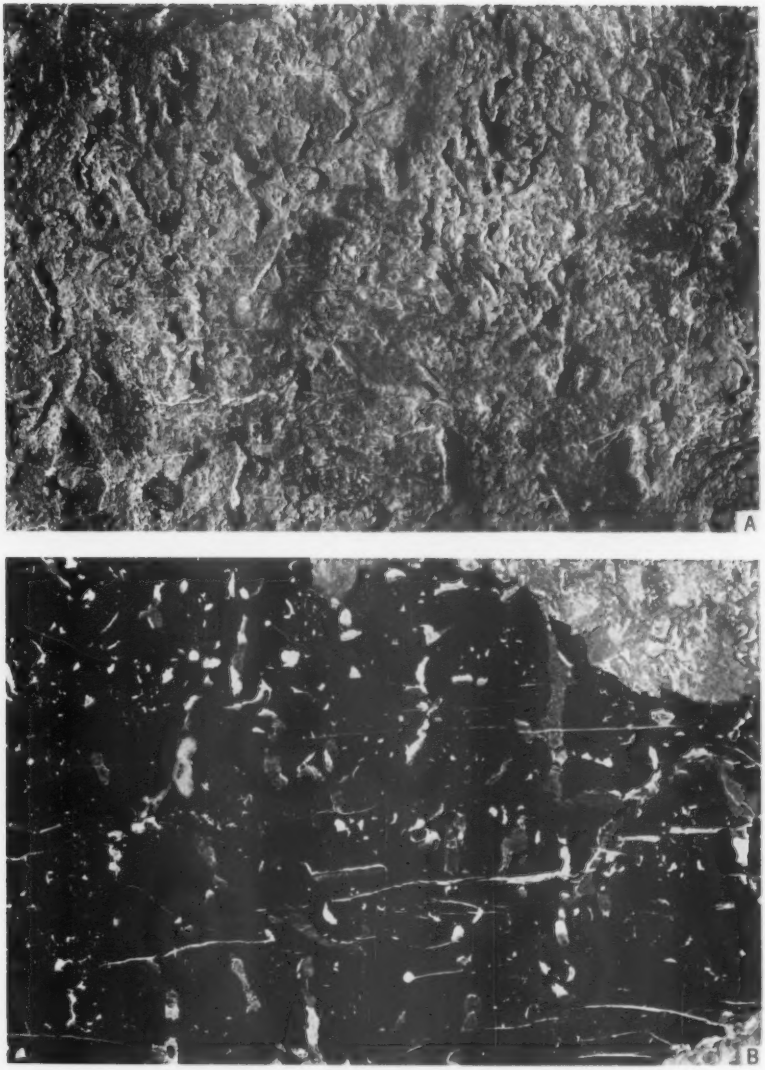
E(× 6.7). Very thin section (crossed nicols) in vertical orientation showing transitional contact from dark microgranular matrix to fine-grained sparry dolomite of supposed interstitial gas accretion site. The circular objects are microfossils, and a little vug-dolomite shows at lower left.

PLATE 1



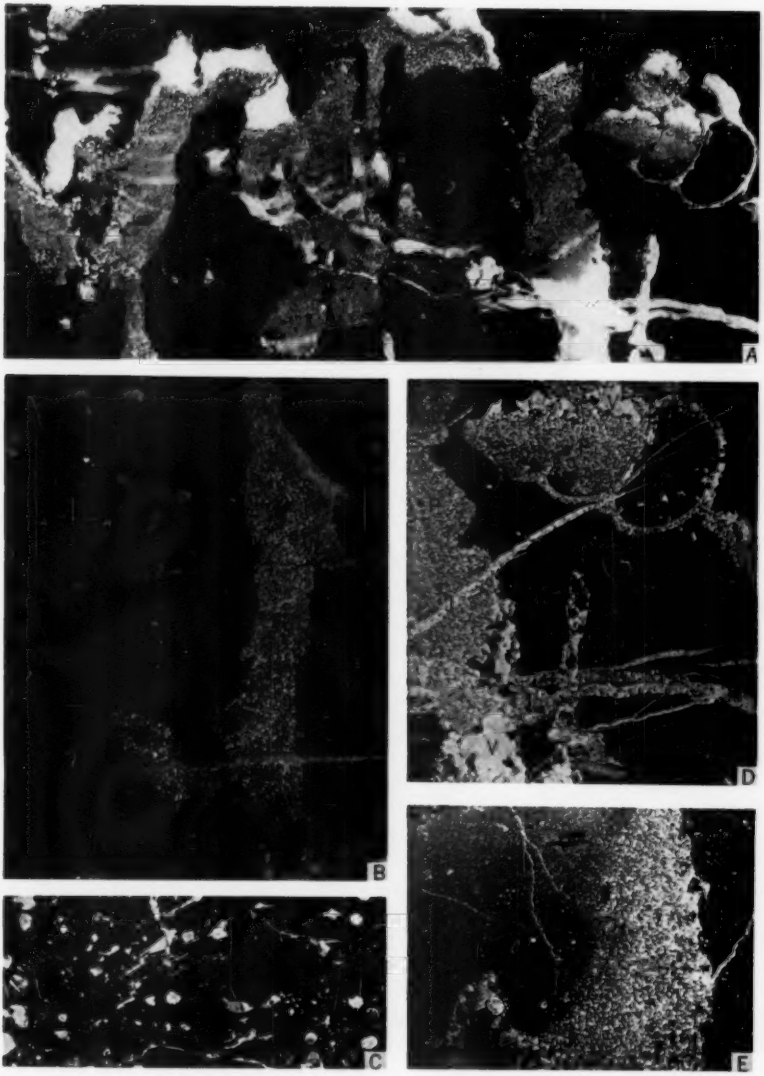
Gas evasion marks on Devonian and modern sediments

PLATE 2



Probable former gas pockets in a Middle Devonian black limestone

PLATE 3



Detail of same probable gas accretion sites as illustrated in plate 2

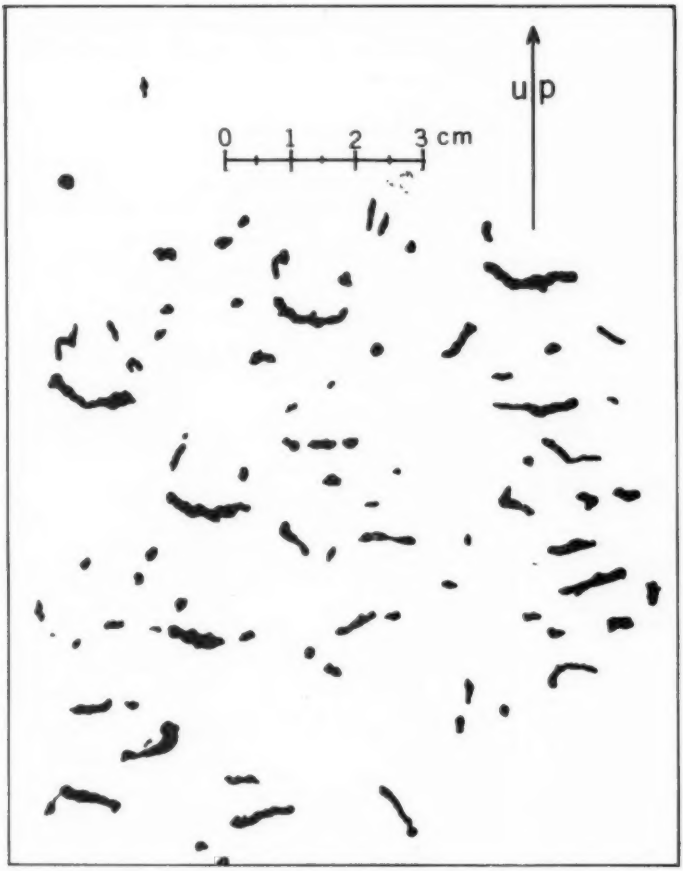


Fig. 1. Tracing of cavities created against inner surface of a sealed plastic core-liner by gathering and moving gas in aragonite mud from Station *It*[1], middle of Bahama Banks west of Andros Island.

axes range in orientation from horizontal to vertical, and they may shift course abruptly from horizontal to vertical (pl. 2B, upper right quarter). Some contract and expand. They show a wavering tendency to taper upward and to be capped by clear, sparry dolomite or calcite; and these vug-like spaces commonly have flat floors which parallel the bedding and imply their very early origin as cavities within a semifluid matrix.

The sites of coarse, clear, vug-filling dolomite or calcite within the markings are locally intersected by equally clear but finer grained and commonly

[1] The carbonate mineralogy was confirmed by x-ray powder pattern records prepared by Mary Mrose, U. S. Geological Survey.

fibrous vein-calcite of subhorizontal to inclined tension joints, with slight to no perceptible offset and thus without involving a significant shear component in the jointing process. Where the vein calcite intersects earlier carbonate minerals it can be seen that they are distinct and that the vein calcite is the younger (pl. 3B, 3D). No pyrite is found within the vein calcite, and only a little in the vug-filling dolomite or calcite, but it is commonly concentrated just below or at the lower edge of the vugs. Generation of the pyrite, therefore, seems mainly to have preceded the vug-filling but partly to have overlapped it.

The contact relations of the pyrite with, and its occasional inclusion within, the early sparry dolomite of the markings favors their near contemporaneity. The irregularities and gradational margins of the markings, and the remnants of unaltered matrix within them, reinforce misgivings about their being simple burrows in the usual sense.

Now we are ready to reconstruct the post-depositional history of this limestone, and to interpret the pattern found in it. The abundance of pyrite in the subcylindrical markings—and the relations between pyrite, associated sparry dolomite, and calcitic matrix rock—suggests the early diagenetic concentration within the boundaries of these markings of bacterially generated hydrogen sulfide gas. Indications of hydrogen sulfide, and the preservation of the organic matter that gives a black color to the limestone, imply anaerobic conditions; while the association of calcite and pyrite implies a pH between 7 and 8 at the time the pyrite formed (Krumbein and Garrels, 1952, p. 19, fig. 8). Probably as a consequence of the formation of the pyrite (or antecedent melnikovite, Lepp, 1957), however, the pH increased; and, either then or later, recrystallization to dolomite occurred within the markings, reducing the original matrix material there to interstitial and insular remnants. It is, to be sure, likely that gas accretion or growth of gas generating bacteria was to some degree facilitated by the presence of disturbed zones in the sediment, and that the structures described may to some extent follow or reflect earlier pathways through it. The preexistence of such disturbed zones, however, is probably not crucial to the development of markings of this general nature.

The paragenetic relations described suggest the following sequence of interrelated main events: (1) deposition of calcium carbonate mud, including minute testaceous organisms (pl. 3D-E) and a high content of organic matter; (2) generation of H_2S by sulfate reducing bacteria within the sediment, creating a reducing interstitial environment slightly above neutral pH, inhibiting organic consumption of the carbon, and leading to the local concentration of gas in interstitial openings and gas generated larger cavities which jointly reflect or created a pattern of burrow-like passageways; (3) nearly concomitant mobilization of iron within the sediments and reaction with H_2S to form pyrite, very likely accompanied by rising pH; (4) a related increase of carbonate ion and CO_2 pressure, with the result that previously dissolved matrix recrystallized as fine-grained dolomite within the limits of the markings described; (5) in continuation of this process, and partly as a result of continued compaction and expulsion of pore water, filling of larger open spaces with clear coarse-grained dolomite and locally calcite; (6) completion of lithification and eventual transition from the diagenetic to the metamorphic realm, followed by

lateral compression which resulted in vertical tensional stresses, subhorizontal to oblique jointing, and joint filling by sparry fibrogranular calcite; (7) exposure to the atmosphere, weathering, and the local formation of limonite-after-pyrite.

PITFALLS

Bubble imprints are easily confused with raindrop imprints. Unless, therefore, the origin of a particular pitted surface is strongly indicated by specific features, it is better not to apply genetic terms to it. Raindrop and other impact prints should show raised rims, which the marks of resting bubbles generally lack, but which the marks of bursting bubbles may have, and which gas craters generally have. The sides of hail imprints are typically steep to overhung (Lyell, 1851, p. 243; Twenhofel, 1921, p. 363, fig. 3). If pitted surfaces are suspected to be the work of rising gas, it is well also to look for disturbance of the sediment through which such gas might have risen. Other differences were cited on a preceding page.

Tracks of drifting bubbles may be mistaken for the trails of "worms" or other soft bodied organisms, especially where they curve about (Twenhofel, 1921, fig. 6). The tracks of wind-blown bubbles, however, are commonly distinguished by a marked approach to rectilinearity and a high degree of parallelism (Reineck, 1954, figs. 1-6). Certain types of rill markings (including solution) also resemble composite bubble or foam tracks except in their massed patterns and abrupt discontinuities; for instance, in 1957 Reineck (oral communication) believed that figures 8 and 9 of his 1954 paper represent a form of swash mark rather than foam streaks, as he originally interpreted them.

James (1884, p. 128-129) long ago realized the possibility of mistaking the marks of gas bubbles for organic markings when he attributed such an origin to two of Walcott's "algal species" from the Utica shale; although, in fact, neither Walcott's nor James' illustrations provide much basis for identification. More recently Alf (1959, fig. 2) has described and illustrated as "jellyfish-like impressions" structures which are probably gas blisters or their inverse (*cf.* pl. 1B; also McKee, 1945, pl. 7d). These specimens are from red shale in the Bass formation of the Precambrian Grand Canyon series; they are associated with a more problematical blister-like structure (Alf, fig. 1) which is preserved on the (under?) surface of a fine-grained slabby quartzite but which is morphologically reminiscent of a multiple gas blister (*cf.* pl. 1A; also Maxson, 1940, fig. 1, upper half)[1]. A thin slab of quartzite in the Nankoweap group of the same Grand Canyon series has yielded still another problematicum that has been cited and named as a "supposed jellyfish" (Bassler, 1941), as an unequivocal marine "jellyfish medusa" (Van Gundy, 1951), and as an unquestioned "protomedusid" (Harrington and Moore, 1956). Although Bassler considered "the best proof" of its organic origin to be its sedimentary relationships, study of this marking reveals that it is probably on the under (rather than the upper) surface of the slab and that (as Van Gundy recognized) the slab is not ripple marked or finely cross-laminated as a retouched

[1] Since this was written I have, through the courtesy of Prof. Alf, had opportunity to examine these or similar specimens; in my judgment all resemble gas blisters or their impressions.

photograph indicates (Bassler, 1941, pl. 64, fig. 3). The "jellyfish" seems to be the reverse imprint of a subradial fracture system of unknown origin in an underlying shale, bits of which are preserved in the fractures. Its general pattern is reminiscent of the ruptured gas domes illustrated by Schäfer (1954, figs. 10-12), but it remains to be demonstrated whether load compression of of such structures could produce sufficient lateral thrust to account for the lobate periphery.

Finally, gas cracking of sediments such as that illustrated by Schäfer (1954), and produced underwater, might even be mistaken for dessication cracking due to subaerial exposure or for brecciation due to wave action in water much shallower than its actual site of formation.

ACKNOWLEDGMENTS

Discussions with W. H. Bradley, Robert George Schmidt, and W. S. White were helpful in developing and clarifying the ideas expressed above. This paper has also had the benefit of critical reading by H. S. Ladd, K. E. Lohman, and W. A. Oliver, and the bibliographic assistance of Mrs. Ruth Gove. I am particularly indebted, moreover, to Mary Mrose and Schmidt for resolving a vexing mineralogical side-issue that arose in connection with this study when calcium formate showed up as a synthetic product in formic acid residues of the limestone illustrated on plates 2 and 3.

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THE BOUNDARY BETWEEN ROCKS OF CARLILE AND NIOBRARA AGE IN SAN JUAN BASIN, NEW MEXICO AND COLORADO*

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ABSTRACT. In the northern part of the San Juan Basin, New Mexico and Colorado, an unconformity at the base of beds of Niobrara age cuts out 300 to 400 feet of beds of late Carlile age that in the southern part of the Basin are represented by beds from the top of the Juana Lopez sandstone member of Rankin (1944) at least up to the top of the Gallego sandstone member of the Gallup sandstone. The Juana Lopez sandstone member of the Carlile shale and the Tres Hermanos sandstone member in the lower part of the Mancos shale are shallow water marine sandstones, the tops of which are nearly time equivalent over very extensive areas. The Gallego sandstone member of the Gallup sandstone of the Mesaverde group may be similar to them. The stratigraphic relations suggest that in a belt 30 to 75 miles wide between the northern and southern parts of the Basin downflexing toward the south occurred during the time of deposition.

INTRODUCTION

The rock units discussed in this paper were laid down in a marginal area of the Late Cretaceous sea, in northwestern New Mexico (fig. 1). The rocks were later warped downward to form the San Juan Basin, a structural depression about 100 miles wide from east to west and about 150 miles from north to south in northwestern New Mexico and southwestern Colorado. The rock units described are exposed along the western and eastern margins of the Basin. The diagrammatic sections shown (figs. 2 and 3) trend about N 15° E. and are nearly normal to the trend of the Cretaceous strand lines, which was about N 60° W.

At the northeastern (seaward) ends of the diagrammatic sections, the Upper Cretaceous rocks are divided into the Dakota sandstone at the base; the overlying Mancos shale, with named members in the northeastern part of the Basin (Dane, 1946) and the Mesaverde group, of which only the basal part of the Point Lookout sandstone is shown on the sections. Southward along the lines of sections, the marine units of the Mancos shale are invaded by successively lower tongues of littoral sandstone and by lagoonal, swamp, and continental deposits that comprise the named units of the Mesaverde group of the southern part of the Basin (Beaumont, Dane, and Sears, 1956).

This summary of available published outcrop information (particularly Pike, 1947; Sears, 1934; Hunt, 1936; Dane, 1946) as modified by the results of some recent stratigraphic and paleontologic studies (Dane, Wanek and Reeside, 1957, and Dane, Bachman and Reeside, 1957) and supplemented by some still more recent field studies in the spring and fall of 1959 by the writer and W. A. Cobban, develops a somewhat different interpretation of the relationships of the Mancos shale and the lower formations of the Mesaverde group than has hitherto appeared, particularly with regard to the boundary between rocks of Carlile and Niobrara age. This interpretation has possible significance in clarifying previously published accounts of the Upper Cretaceous surface stratigraphy of the northern part of the Basin.

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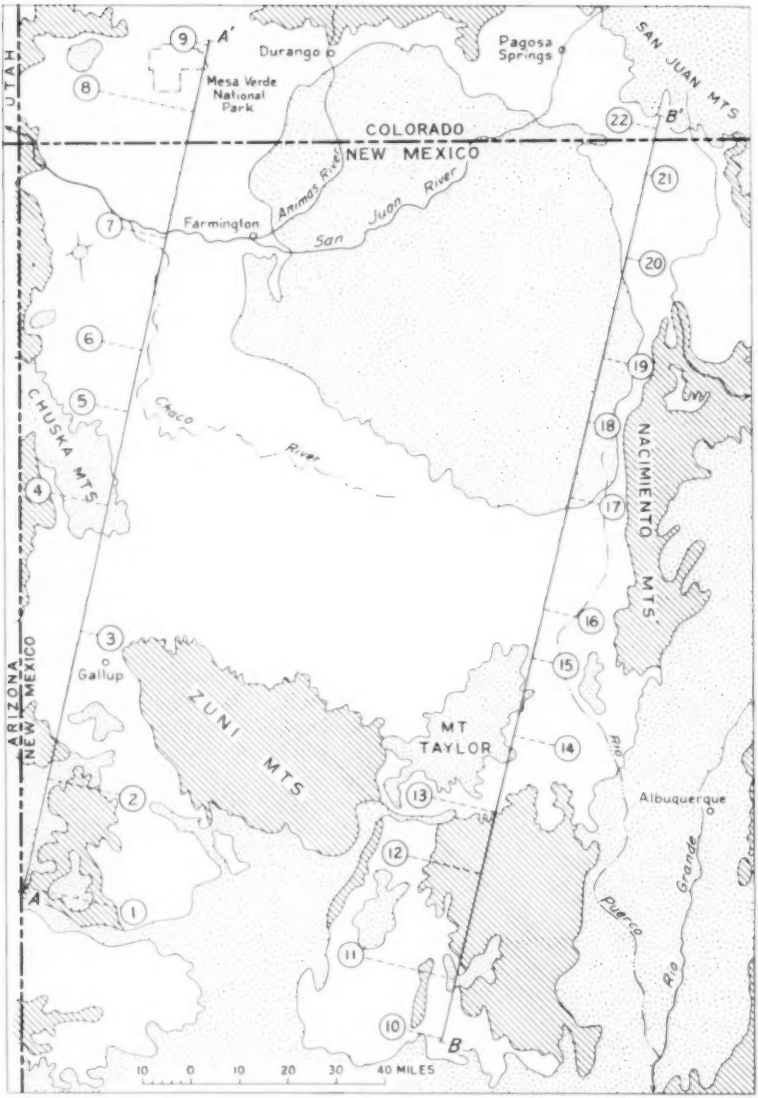

Fig. 1. Index map of part of northwestern New Mexico and southwestern Colorado showing locality numbers projected into lines of sections A-A′ (fig. 2.) and B-B′ (fig. 3.). Tertiary and Quaternary rocks and surficial deposits (stippled), Jurassic and older rocks (lined) and Cretaceous rocks (unpatterned).

Because of the brevity of this paper, it proposes no changes in the existing status of nomenclature although some are obviously needed.

THE DAKOTA SANDSTONE, TRES HERMANOS SANDSTONE MEMBER OF MANCOS SHALE, GRANEROS SHALE, AND GREENHORN LIMESTONE

The deposits associated with the earliest transgression of the Late Cretaceous sea in this region include carbonaceous shale, coal, and a varied assemblage of shales, mudstones, sandstones and conglomerates. They are generally of lagoonal and continental origin in the lower part, and marine or at least trending toward marine in origin in the upper part, at least in so far as this can be inferred from the increased regularity of bedding of sandstone, the abundance and character of ripple marks, the abundance of various borings and trails of invertebrates and the character of the associated shales. These rocks, generally ranging from 120 to 250 feet or more in thickness have been included in the Dakota sandstone by many writers (Dane, 1946; Renick, 1931; Reeside, 1924, p. 9, and others). Qualifications are expressed in many of the papers as to the propriety of the usage, however, because of uncertainty of correlation with the far removed type locality. This so-called Dakota in the northern and central parts of both east and west sides of the Basin is succeeded by a widespread unit, dominantly dark marine shale with interbedded bentonite, that has commonly been referred to as the Graneros shale, although it is in part at least of the age of the Greenhorn limestone of Kansas. The Graneros in turn is succeeded over most of the Basin by a unit of white weathering thin-bedded gray limestone with a maximum thickness of about 50 feet and containing the fauna of the upper part (Bridge Creek limestone member) of the Greenhorn limestone of Kansas. A thin remnant of Greenhorn is known to be present as far south as Mesa Prieta (east of locality 15) on the east side of the Basin and at least as far south as Toadlena (locality 5) on the west side of the Basin. It is believed to lie close to a single time horizon throughout its extent in the San Juan Basin, because of 1) its consistent position immediately succeeding a widely correlatable distinctive sequence of bentonite beds, to be described elsewhere; 2) its consistent position with respect to faunal zones so far as can be determined; and 3) the consistency of interval to key horizons above and below. Therefore, it has been used as a datum in drawing the sections.

East of Mt. Taylor and on the south side of the Basin the name Dakota has been applied to only part of what has elsewhere been called Dakota—a dominantly sandstone and conglomerate unit 15 to 50 feet thick but locally missing, which in the vicinity of Acoma (south of locality 13) contains marine fossils of the oldest Upper Cretaceous faunal zone (Dane, 1959, p. 90). It is overlain by 150 to 200 feet of marine shale, containing at least three sandstone beds, of which the most persistent, thickest and extensive is the Tres Hermanos sandstone member of the Mancos shale. This sandstone is 20 to 40 feet thick and equivalent in age to the middle or lower part of the Greenhorn limestone of Kansas. It extends northward nearly to the vicinity of Senorito (locality 17) where its thinning wedge edge lies about halfway between the Greenhorn lime-

stone and the top of the Dakota. In the upper part of the Tres Hermanos or in the immediately overlying shales, a broad undescribed species of *Gryphaea* is abundantly and extensively present. This species is present in the lower part

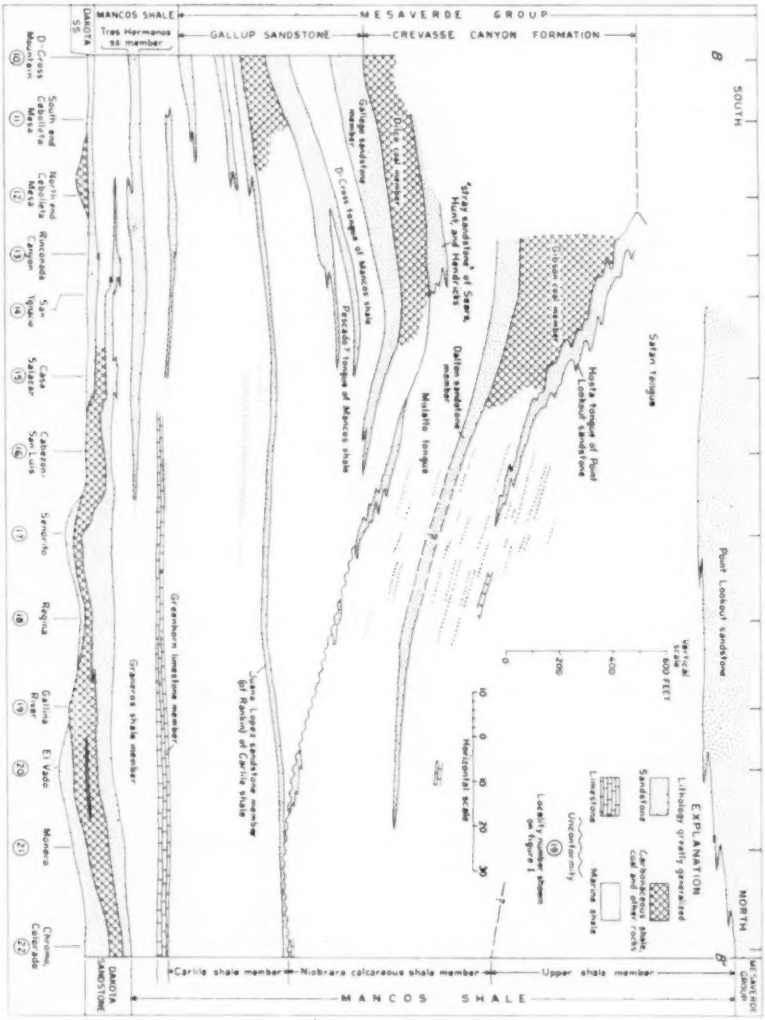

Fig. 2. Diagrammatic interpretation of the intertonguing relationships of the Dakota sandstone, Mancos shale and the lower formations of the Mesaverde group on the west side of the San Juan Basin from Atarque, Valencia County, New Mexico to Point Lookout, Mesa Verde National Park, Colorado.

(Lincoln limestone and Hartland shale members) of the Greenhorn limestone of Kansas. (W. A. Cobban, written communication).

South of the vicinity of Mt. Taylor, the Tres Hermanos sandstone member is thicker, coarse grained, more highly cross-bedded, and less fossiliferous. A similar change from marine to continental beds in the underlying Dakota indicates a southerly source, as does the termination northward of these units east of Mesa Prieta (northeast of locality 15). However, on the east side of the San Juan Basin the much coarser grained, thicker and more continental material of the so-called Dakota suggests that it may there have had a source to the east and that, before and during Dakota deposition, uplift and erosion occurred in the area of the Nacimiento Mountains or San Juan Mountains to the northeast, where probable overlap of the Dakota on Precambrian rocks has been reported by W. R. Muehlberger (written communication).

The consistent interval between the Tres Hermanos sandsone member and the Greenhorn limestone and its uniform position with respect to faunal horizons and to bentonite beds above and below in so far as has been determined suggest that, despite its great extent, it departs but little from time horizons.

Although the correlation of the Tres Hermanos across the southern part of the San Juan Basin has not been established, it is believed to be the same unit as the Twowells sandstone lentil of Pike and a *Gryphaea* bearing sandstone about 80 feet above the top of the Dakota sandstone at Gallup (Dane and Bachman, 1957). This Tres Hermanos equivalent is more closely approached by the Dakota northward so that they become in effect one unit near Shiprock (locality 7). The Graneros shale member of the Mancos shale above this combined unit is thus thinner in the northwestern part of the Basin than it is in the northeastern part, where it continues downward below the horizon of the Tres Hermanos sandstone member.

BEDS OF CARLILE AGE IN THE SOUTHERN PART OF SAN JUAN BASIN

The Greenhorn limestone is suceeded by about 450 to 500 feet of very dark shales, containing characteristically large yellow weathering septarian carbonate concretions and carrying in the lower part the *Collignoniceras woollgari* fauna of the lower part of the Carlile shale (Cobban and Reeside, 1952, p. 1018). This lithologic unit forms the lower and principal part of what has been called, in the northeastern part of the San Juan Basin, the Carlile sandy shale member of the Mancos shale (Dane, 1946).

The upper part is represented by an extremely widespread sandstone unit, 10 to 40 feet thick, called the Juana Lopez sandstone member (of the Carlile shale) by Rankin (1944). This, with its equivalents, extends far into Arizona, Utah, and Colorado. This sandstone is more generally called by commercial geologists the Sanostee or Sanastee[1] member of the Carlile shale or of the Mancos shale (Momper and Tyrrell, 1957, p. 23; Smith, C. T., and others 1951, p. 45-46, 79). It is characteristically hard, brown, calcareous, thin-bedded, platy or slabby fine grained sandstone, containing abundant oyster

[1] Should be spelled Sanostee following the spelling of the name of the Navajo Indian Trading Post, the type locality, as shown on U. S. Geological Survey topographic map Sanostee quadrangle, (1953) 15 minute series.

shells, *Inoceramus* fragments and shell prisms, shark teeth, fishbones, and phosphatic nodules. Its lithology, its wide extent at a position generally about the same distance above the Greenhorn limestone and top of the Tres Hermanos sandstone, and its approximately uniform position in the lower part of the paleontologic zone of *Prionocyclus wyomingensis* (Cobban and Reeside, 1952, p. 1018) suggest that it is essentially a time equivalent unit throughout its ex-

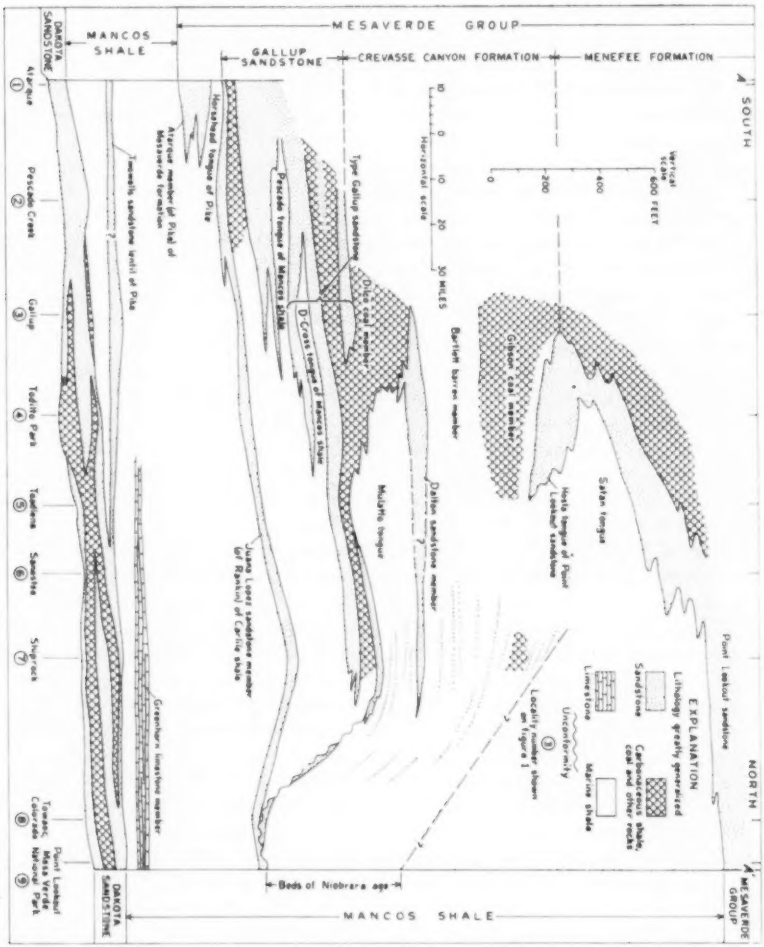

Fig. 3. Diagrammatic interpretation of the intertonguing relationships of the Dakota sandstone, Mancos shale, and the lower formations of the Mesaverde group on the east side of the San Juan Basin from D-Cross Mountain, Socorro County, New Mexico, to Chromo, Archuleta County, Colorado.

tent in the San Juan Basin and probably over a much larger region to the north and east.

To the south, temporal and lithologic equivalents of Rankin's Juana Lopez are included in the basal part of the Gallup sandstone as delimited by Pike in the Atarque area (locality 1), and in the middle to lower part of the Gallup as used by Pike and by Dane and others in the D-Cross Mountain area (locality 10).

In the southern part of the region here considered, the Juana Lopez sandstone member and its equivalents are overlain in sequence by an unnamed coal and carbonaceous unit, by tongues of the Gallup sandstone, by tongues of the Mancos shale (the D-Cross and Pescado tongues) and then by the Gallego sandstone member of the Gallup sandstone (Dane, Wanek, and Reeside, 1957) and younger beds.

Until recently there has been little specific information on the age of the Gallup sandstone. In 1957 the fauna of the D-Cross tongue and the basal part of the overlying Gallego sandstone member was hurriedly collected under adverse conditions by the late J. B. Reeside, Jr. and determined to be of latest Carlile age (Dane, Wanek and Reeside, 1957, p. 192-195). Subsequent stratigraphic studies (Dane, Bachman and Reeside, 1957) extended the D-Cross tongue and the supposed latest Carlile faunal zone to the southern part of the Mt. Taylor coal field and into the lower part of the Gallup sandstone as defined at the type locality. In the Spring of 1959 more extensive collections from the D-Cross tongue were made by W. A. Cobban in company with G. O. Bachman and the writer, and the entire fauna was carefully reviewed. The fauna of the D-Cross tongue was placed as about middle Carlile, although the presence of several species not previously known below beds of Niobrara age suggests that it may be slightly younger than the fauna of the Juana Lopez sandstone member of Rankin. Cobban, however, has collected in the beds just below the Gallup sandstone west of Shiprock a species of *Inoceramus* that he regards as of latest Carlile age. So far as known, species diagnostic of age have not been found in the upper part of the Gallego sandstone member of the southern part of the Basin or of the upper part of the Gallup on the western side. In the fall of 1959, however, D. L. Wyant collected a single specimen of the ammonite genus *Barroisiceras* from the upper part of the Gallego sandstone south of San Luis (locality 16). This is a genus regarded by Cobban as diagnostic of the basal part of the Niobrara formation (basal part of the Fort Hays limestone member) of the Pueblo, Colorado, region. Some stratigraphic rise seaward of the Gallego thus seems likely, but it has not yet been conclusively demonstrated. The Gallego resembles in its general relationships, habit, extent, and comparative uniformity of interval above known time equivalent horizons the Tres Hermanos and Juana Lopez sandstone members. Richly titaniferous sandstone deposits at several localities at the top of the Gallup (Murphy, 1956) indicate a long continued winnowing action not inconsistent with the notion that the top is time equivalent over etxensive areas.

A more conspicuous stratigraphic and chronologic rise of the Gallup seaward has, however, previously been presumed, chiefly on the basis of occurrences of *Inoceramus "umbonatus"* (correctly *Inoceramus involutus*) a species

of middle Niobrara age, in beds previously included in the Gallup in the vicinity of the San Juan River (Reeside, 1924, pl. 3; Pike, 1947, p. 50). The occurrences are now known to be only in beds that are not equivalent to and should be excluded from the Gallup sandstone.

UNCONFORMITY AT BASE OF BEDS OF NIOBRARA AGE IN THE NORTHERN PART OF SAN JUAN BASIN

The name Gallup sandstone was extended, by Pike (1947, p. 28, 29) and later by Beaumont, Dane, and Sears (1956, p. 2156), to the tripartite unit in the vicinity of Shiprock (locality 7), part or all of which had been named by Reeside (1924, p. 9) the Tocito sandstone lentil of the Mancos shale, although Pike recognized that the upper sandstone bed of the unit was quite likely an equivalent of the "stray" sandstone of Sears, Hunt, and Hendricks (1941, p. 113 and pl. 26). The medial soft unit, a carbonaceous shale and sandstone unit presumably equivalent to the Dilco coal member, is 25 to 55 feet thick in the vicinity of Shiprock. It grades laterally northward into beds containing *Ostrea sannionis* and ultimately into marine shale. The lower sandstone of the tripartite unit is bedded, light brown weathering fine to medium grained sandstone 45 to 60 feet thick. It thins and disappears just south of the San Juan River. The upper sandstone of the tripartite unit is a coarse-grained, cross-bedded, clean quartz sandstone 20 to 40 feet thick, locally with small pebbles, chiefly siliceous. It continues northward beyond the San Juan River as a well defined ledge of coarse-grained quartz sandstone. This upper sandstone, markedly glauconitic in part, locally contains *Inoceramus involutus* (*Inoceramus umbonatus* of authors) of middle Niobrara age (Cobban and Reeside, 1952, p. 1019), and is succeeded by sandy shales that carry within a few feet of the base the very large flat *Inoceramus* of middle to late Niobrara age that has for many years been called *Inoceramus* (*Haploscapha*) grandis.

Northward from the San Juan River the glauconitic sandstone cuts sharply downward across the underlying beds. At a locality 3 miles south of the Colorado State line coarse grained glauconitic sandstone rests directly on shales of Carlile age within 60 feet of Rankin's Juana Lopez sandstone member. West of Towaoc, Colorado, (locality 8) south of Ute Mountains, coarse grained glauconitic quartz sandstone containing *Inoceramus deformis* of Niobrara age rests on only 15 feet of shale above the Juana Lopez sandstone member. It is of interest to note that the basal few inches of the glauconitic sand contains a few shells of *Ostrea lugubris*, doubtless reworked from the underlying shales in which it is also present. Elsewhere in this area glauconitic sand of Niobrara age apparently rests directly on sandstone of the Juana Lopez and has been included with it (Ekren and Houser, 1958, p. 74-77). At Point Lookout entrance station to Mesaverde National Park, some poor exposures suggest that shales of Niobrara age rest directly on the Juana Lopez sandstone member.

The conclusion seems inescapable that beds of the *Inoceramus deformis* zone cut down unconformably northward and that 300 feet or more of beds present south of San Juan River are missing just south of Ute Mountains about 30 miles to the north.

A closely comparable situation exists along the east side of the Basin. Twenty-five miles northeast of the termination of the Gallup sandstone and 7 miles south of Senorito (locality 17) a thin glauconitic sandstone containing *Inoceramus grandis* of authors rests on 450 feet of shale probably all of Carlile age above the Juana Lopez sandstone member. At El Vado (locality 20), however, a coarse glauconitic sand containing *Inoceramus deformis* rests on only 30 feet of shale above the Juana Lopez and a few miles farther north (locality 21) 30 feet of thin-bedded shale containing *Inoceramus deformis* rests directly on the Juana Lopez and is succeeded by calcareous shales containing *Inoceramus grandis* of authors and *Ostrea congesta* of middle Niobrara age. Exposures of this stratigraphic interval between localities 17 and 20 are poor and incompletely studied, but the unconformable relationship seems evident. The truncation of beds is distributed over a south-north distance of perhaps 75 miles as compared with 30 miles north of Shiprock (locality 7).

It is suggested that this unconformity is the result of limited marine erosion in a general area of non-deposition. It is noteworthy that the Gallup sandstone wedges out a short distance south of the point where, both on the west and east sides of the Basin, thinning of the section is first evident. It is also noteworthy that the occurrence of markedly glauconitic sands at the base of beds of Niobrara age is apparently limited on both sides of the Basin to the belt of unconformable relationship. More detailed studies are needed fully to establish and to interpret the relationships at this horizon.

Cloud (1955) has summarized the physical limits of glauconite formation, suggesting that its formation is favored by such factors as normal marine waters, depths in the upper part of the 10 to 400 fathom interval, the availability of micaceous minerals or bottom muds of high iron content, slightly reducing conditions, and generally slight sedimentary influx or bypassing of the finer fractions. All of these conditions seem likely to have developed with subsidence, transgression, and a marine unconformity at this horizon and in this part of the basin.

It may be significant that the availability of iron in the sediments near this horizon is shown by the numerous pebbles of mafic igneous rocks in sands, probably in the basal beds of the Niobrara, reported in the Hogback and Rattlesnake areas near Shiprock (locality 7) (Beaumont, 1957, p. 118).

The stratigraphic relations in the San Juan Basin south of the belt of loss of section by non-deposition and marine truncation present a general picture of the deposition of very widespread, approximately time parallel sandstones, almost certainly so for the Tres Hermanos and Juana Lopez sandstone members, and quite likely for the Gallego sandstone member and underlying tongues of the Gallup sandstone. The belt of loss of section of rocks of late Carlile age that extends in a west northwesterly direction across the northern part of the San Juan Basin seems therefore to be most readily explained as a result of a relatively abrupt hinge warping of several hundred feet downward to the south along this belt during the time when the post-Juana Lopez beds of the southern part of the Basin were being deposited. The belt of downwarping bordered an

extensive relatively high platform to the north, and an area of generally uniform subsidence to the south, but a suggestion that extremely broad and gentle downtilting of the southern part of the Basin also occurred at times during transgressive phases is given by the southward thickening of the D-Cross tongue and of the shale unit below the Tres Hermanos sandstone member.

Evidence of comparable structural movements in eastern Colorado during deposition in the early part of Late Cretaceous time was presented many years ago by Johnson and Aurand (1929, p. 850-853) and the evidence for an unconformity at the base of the Niobrara formation was summarized by Johnson (1930, p. 789-794). Local thinning within the Carlile in the vicinity of La Junta, in the Arkansas valley of eastern Colorado, has also been attributed to a gradual structural rise during deposition (Dane, Pierce and Reeside, 1937, p. 219-220). The beds immediately above the unconformity in the northern part of the San Juan Basin correspond in age with those of a horizon some feet above the base of the Niobrara of the Arkansas valley. This suggests either that the structural movements ended earlier in the Arkansas valley area or that a longer period of submarine erosion or non-deposition intervened between the termination of structural movement and the beginning of deposition in the San Juan Basin.

The relationships described show that beds from the top of the Juana Lopez sandstone member at least to the top of the Gallego sandstone member have no temporal or lithologic equivalents in the northern part of the San Juan Basin, and the D-Cross and Pescado tongues of the Mancos shale thus have no equivalents at the type locality of the Mancos. They obviously connected to the east with marine shales that have likewise been included in the Mancos shale and were deposited in what was probably the main body of the Late Cretaceous sea. The source of the sandstones of the Mesaverde formation to the southwest, and a west northwest trend of their terminal edges have been amply demonstrated, but the present San Juan Basin lay on a shelf lateral to the main Late Cretaceous seaway. It is believed that this marginal shelf situation prevailed throughout the early part of Late Cretaceous time and that tongues of shale, like that below the Tres Hermanos sandstone member for example, spread laterally west northwestward across this shelf. The marginal edges of some of these shale tongues had a quite different trend from the marginal edges of the sandstone units of the Mesaverde group.

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OXIDATION OF PYRITE BY IRON SULFATE SOLUTIONS*

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ABSTRACT. The rate of oxidation of pyrite specimens from three localities was measured in acid iron sulfate solutions. The rate of reduction of ferric ion in these solutions is considered to be a measure of the rate of oxidation of the pyrite. The overall oxidation reaction, in the range 100 percent to 0.1 percent $m_{Fe^{+++}}$ is in accord with the classical reaction:

$$FeS_2 + 8H_2O + 14Fe^{+++} = 15Fe^{++} + 2SO_4^{--} + 16H^+.$$

The average rate of reduction from 100 percent $m_{Fe^{+++}}$ to 50 percent $m_{Fe^{+++}}$ is constant for pyrite from a given locality, and differences in the average rates between specimens from different localities could not be related to minor compositional variations of the solutions or of the pyrites. The pH did not affect rate in the tested range, pH 0 to pH 2.

The instantaneous rate of reduction of ferric ion diminishes with decrease in the ferric-ferrous ratio; it is postulated that instantaneous rate is controlled by differential absorption of ferric and ferrous ions on the pyrite surface, and is proportional to the fraction of pyrite surface occupied by ferric ion.

INTRODUCTION

The oxidation of pyrite in nature is a complex process, involving both organic and inorganic agents. Temple and Koehler (1954) have reviewed the literature, and have made a critical summary of the major factors to be considered.

This paper is restricted to one aspect of the oxidation problem. Because ferric salts are commonly cited (Lindgren, p. 829) as important agents in pyrite oxidation, we studied the action of acid ferric sulfate solutions on pyrite samples from three localities. The solutions were kept acid so that no solids would be produced as oxidation products. The rate of oxidation was determined as a function of pH and ferric sulfate concentration, and the effect of minor amounts of various impurities in solution was assessed.

In essence the experiments consist of permitting clean ground samples of pyrite, sized and weighed so as to ensure a nearly constant surface area, to react with known concentrations of acid ferric sulfate solutions at constant temperature and stirring rate. The concentration of ferric iron was continuously monitored.

EXPERIMENTAL MATERIALS AND PROCEDURE

The three pyrite specimens used in the experiments were obtained from the Harvard Museum collection, and consisted of large crystals from Bingham, Utah; smaller crystals, associated with chlorite, from Chester, Vermont; and fine grained pyrite associated with chalcopyrite, from Rio Tinto, Huelva, Spain. Accessory elements found in the three samples are indicated in table 1.

The pyrite was crushed, and the size fraction between 100 and 200 mesh used for the experiments. Shortly before each experiment the pyrite sample was rinsed with 6N H_2SO_4, and with distilled water, and was dried with acetone.

Each initial solution of acid ferric sulfate was obtained by preparing an acid ferrous sulfate solution of known strength and oxidizing it quantitatively

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to ferric by potentiometric titration with cerate or permanganate. The pH was adjusted by addition of H_2SO_4. Atmospheric oxygen and carbon dioxide were excluded by maintaining a nitrogen atmosphere above the solution.

A weighed sample of pyrite was added to the oxidized solution, and a continuous plot of the potential of the solution and elapsed time was recorded automatically. The solution was stirred by means of a teflon coated magnetic bar placed inside the container. The temperature of the experiments was about 33°C. The small changes in temperature (±2°C) during the experiments are probably responsible for part of the scattering of points in the experimental results. Each experiment was allowed to run for from 24 to 100 hours.

TABLE 1

Spectrographic analyses of the pyrite specimens

Jun Ito, analyst

Element	Locality of specimen		
	Bingham, Utah	Rio Tinto, Huelva, Spain	Chester, Vt.
	%	%	%
Cu	0.0X	0.0X	0.0X
Zn	0.X	—	—
Ni	—	—	0.0X
Co	—	0.0X	0.0X
Ag	0.000	0.000	0.000
As	vvs	vvs	—
Ti	—	vvs	0.0X
Mg	tr.	tr.	tr.
Mn	tr.	tr.	tr.
Al	tr.	tr.	tr.
Si	tr.	tr.	tr.

vvs—very, very small
tr.—trace

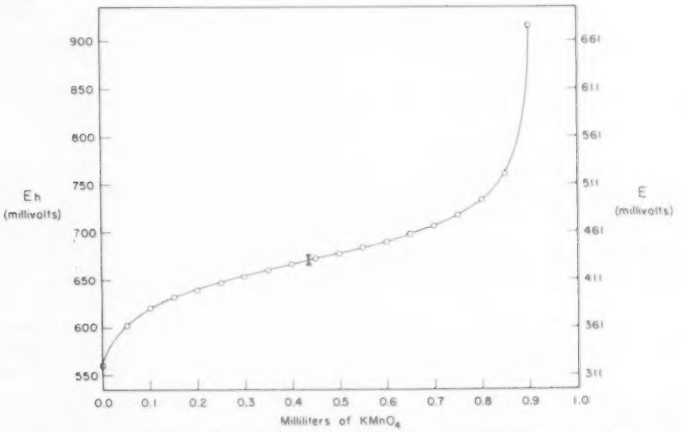

Fig. 1. Potentiometric titration of ferrous sulfate with potassium permanganate. pH 1.55; total dissolved iron ($m^{\circ}_{Fe^{+++}}$) = 1×10^{-3}m/1; $E^{\circ\prime} = 0.671$ volts.

RESULTS

Figure 1 shows a typical curve for the titration of ferrous sulfate by permanganate, prior to reduction by pyrite. From the equivalence point the formal E° ($E^{\circ\prime}$) is obtained. This potential is related to the molality of ferrous and ferric ions as follows:

$$Eh = E^{\circ\prime} + 0.059 \log \frac{m_{Fe^{+++}}}{m_{Fe^{++}}} \qquad (1)$$

The activities of the ions are related by the analogous equation containing E°:

$$Eh = E^\circ + 0.059 \log \frac{a_{Fe^{+++}}}{a_{Fe^{++}}} \qquad (2)$$

Because our interest was in the amount of pyrite oxidized, we used equation (1) almost exclusively. The measured oxidation curves of the various ferrous sulfate solutions agree well with curves calculated from $E^{\circ\prime}$ and the change in $m_{Fe^{+++}}$ and $m_{Fe^{++}}$ expected from addition of increments of oxidizing agent.

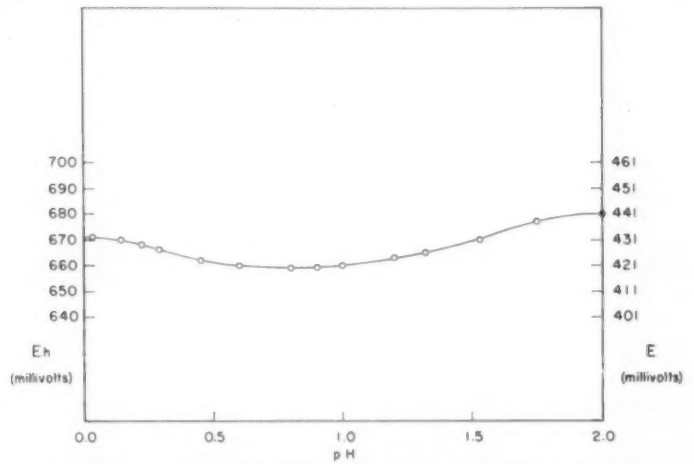

Fig. 2. Change in formal E° ($E^{\circ\prime}$) in acid iron sulfate solutions in the range pH 2 to pH 0.

Figure 2 shows the change in $E^{\circ\prime}$ with pH in the H_2SO_4—$FeSO_4$ solutions used. The small change in $E^{\circ\prime}$ over the large range of ionic strength from pH=2 to pH=0 (the contribution from $FeSO_4$ was always negligible) is surprising, inasmuch as E° and $E^{\circ\prime}$ are related by

$$E^{\circ\prime} - E^\circ = 0.059 \log \gamma_{Fe^{+++}}/\gamma_{Fe^{++}}.$$

The activity coefficients (γ) of the triply charged ferric ion and the doubly charged ferrous ion would be expected to change differentially with increase in ionic strength, and hence $E^{\circ\prime}$ should change strikingly toward lower values at higher ionic strength, that is, lower pH. The ratio of $\gamma_{Fe^{+++}}/\gamma_{Fe^{++}}$ is, however,

nearly constant, for E° is 0.771 volts, and $E^{\circ\prime}$ ranges only from 0.660 to 0.680 volts. The gamma ratio averages about 0.02.

Results of the systematic runs on the reduction of ferric ion with two grams of sized pyrite, in which the time required for 50 percent reduction of the ferric iron was determined, are shown in figures 3, 4, 5, and 6. The coordinates used are those found empirically to be most useful for plotting. The potential versus logarithm of time produces a nearly straight line during reduction to 50 percent $m_{Fe^{+++}}$. The differences between would-be duplicate runs

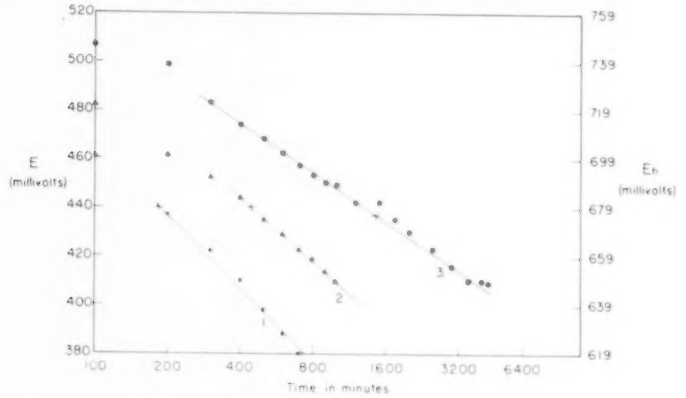

Fig. 3. Rate of reduction of ferric iron in solutions having initial concentrations of 4×10^{-4} $m_{Fe^{+++}}$ by pyrites from three localities. (1) Rio Tinto, Huelva, Spain; (2) Chester, Vermont; (3) Bingham, Utah. Plus marks indicate time to reduction of 50% ferric iron.

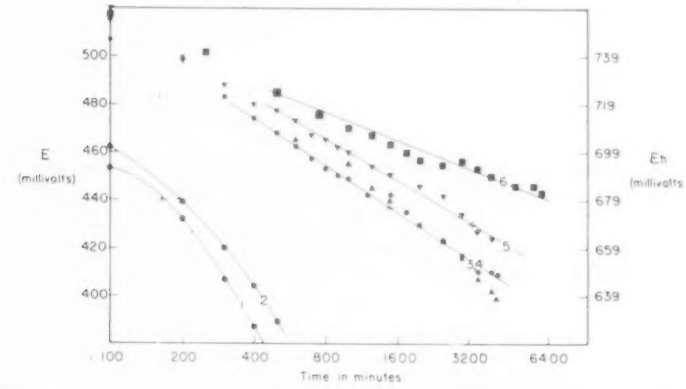

Fig. 4. Rate of reduction of ferric iron by pyrite from Bingham, Utah, in solutions having various initial concentrations of ferric. (1), (2) $m^\circ_{Fe^{+++}} = 4.08 \times 10^{-6}$ m/l; (3), (4) $m^\circ_{Fe^{+++}} = 4.16 \times 10^{-4}$m/l; (5) $m^\circ_{Fe^{+++}} = 1 \times 10^{-3}$m/l; (6) $m^\circ_{Fe^{+++}} = 2.08 \times 10^{-5}$m/l.

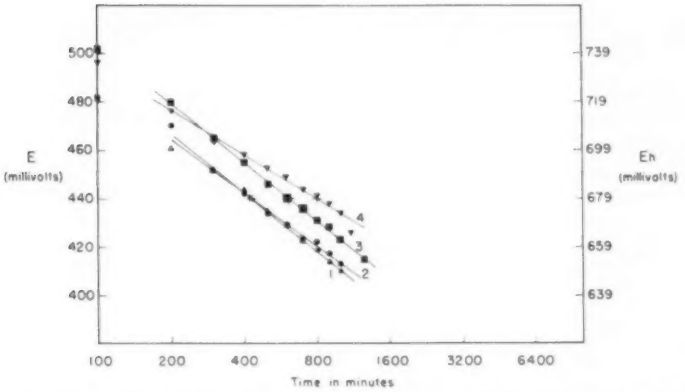

Fig. 5. Rate of reduction of ferric iron by pyrite from Chester, Vermont, in solutions having various initial concentrations of ferric .(1), (2) $m^\circ_{Fe^{+++}} = 4 \times 10^{-4}$m/1; (3), (4) $m^\circ_{Fe^{+++}} = 1 \times 10^{-4}$m/1.

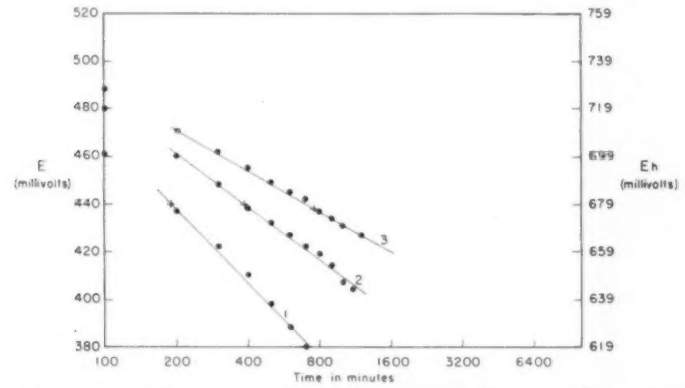

Fig. 6. Rate of reduction of ferric iron by pyrite from Rio Tinto, Huelva, Spain, in solutions having various initial concentrations of ferric. (1) $m^\circ_{Fe^{+++}} = 4 \times 10^{-4}$m/1; (2) $m^\circ_{Fe^{+++}} = 1 \times 10^{-3}$m/1; (3) $m^\circ_{Fe^{+++}} = 2 \times 10^{-3}$m/1.

show the magnitude of the error expected in assigning numbers to the times required for half-reduction of a given sample.

Table 2 shows the results of addition of small amounts of impurities to the solutions, in an attempt to assess the possible catalytic action of typical inorganic constituents of natural waters.

Figure 7 shows results of a typical attempt to achieve equilibrium by adding a large amount of fine ground pyrite (−200 mesh) to a dilute iron sulfate solution. The potential drops rapidly, then becomes irregular at the lower potentials, rising and falling with time. No acceptable equilibrium potential was obtained.

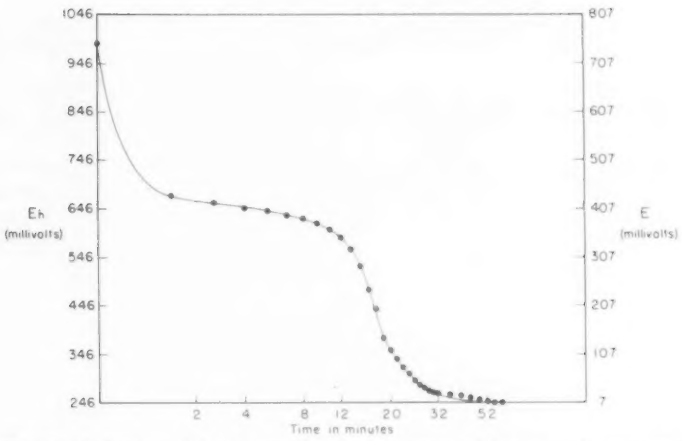

Fig. 7. Reduction of ferric iron by a large amount of fine ground pyrite, $m^{\circ}_{Fe^{+++}} = 2 \times 10^{-4}$m/1; 9.57 gms pyrite.

TABLE 2

The effect of other metal ions in solution on the rate of reduction of Fe^{+++} by pyrite

Pyrite used: Bingham, Utah, 2gms., –100 +200 mesh.
Original Fe^{+++} concentration: 4×10^{-4} molal.

Time to reduce 50% of Fe^{+++} (minutes)	Metal ion	Anion	Concentration (molal)	No. of runs
1240-2000 average 1700	—	—	—	6
1100	VO^{++}	SO_4^{--}	8×10^{-5}	3
1100	VO^{++}	SO_4^{--}	4×10^{-4}	1
1450	VO^{++}	SO_4^{--}	1.2×10^{-5}	2
970	Co^{++}	NO_3^-	2×10^{-4}	3
2000	Co^{++}	NO_3^-	3×10^{-5}	1
2330	Ni^{++}	SO_4^{--}	8×10^{-5}	1
3250	Cu^{++}	SO_4^{--}	8×10^{-5}	1

In several runs, in each of which more than 90 percent of ferric iron was reduced, the solution was retitrated with ceric sulfate. The titrations indicated an increase in dissolved iron of about 7 percent as a result of pyrite oxidation.

DISCUSSION

Variables not affecting pyrite oxidation.—So many variables are involved in a study of oxidation rate that the first step is to eliminate those that show little effect on the process.

A first concern was the possibility that the use of an oxidizing agent to change ferrous iron to ferric iron might influence the ensuing process of reduction of the ferric iron by pyrite. However, runs in which ceric sulfate was used as the oxidant did not differ significantly from those in which potassium

permanganate was used. It was discovered, however, that if an excess of either oxidant was employed, so that the oxidation potential rose temporarily to more than one volt, the behavior of the platinum electrode was affected. It was also discovered, as an interesting sidelight, that the rate of reduction of either ceric sulfate or potassium permanganate by pyrite is very rapid. The typical two gram charge of −100 +200 mesh pyrite will decolorize 250 ml of 0.001 N $KMnO_4$ solution within a few minutes.

Second, pH apparently has little effect on oxidation rate in the range 0 to 2. At higher pH values, where $Fe(OH)_3$ would be expected as an oxidation product, marked effects would undoubtedly occur.

Third, although our experimentation did not systematically explore possible effects, the presence of small concentrations of vanadium, copper, cobalt, or nickel salts (or cerium or manganese as deduced above) had little effect on rate. There is a suggestion of acceleration by vanadium, and retardation by copper, but the effects are probably within our limits of duplication of runs (table 2).

The effects of stirring rate and of anions other than sulfate, either of which might be important, were not investigated.

Oxidation rate as a function of time and ferric iron concentration.—The regular relation between oxidation potential and time, as illustrated by figures 3, 4, 5, and 6, indicates that oxidation of pyrite may be governed by a simple mechanism, except perhaps at low potentials, and the progressive slowing with accompanying lower values points to the concentration of ferric iron as a control.

However, figures 4, 5, and 6 show that the time required for reduction of 50 percent of the ferric iron increases regularly with the original ferric iron concentration ($m°_{Fe^{+++}}$). The average rate of reduction for a given pyrite speci-

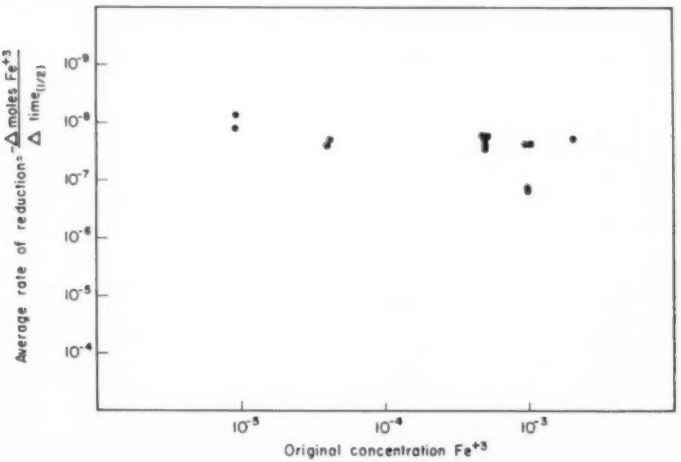

Fig. 8. Average rate of reduction of ferric iron as a function of initial ferric iron concentration, ($m°_{Fe^{+++}}$).

men is independent of total iron content of the solution (fig. 8). On the other hand, for a given experiment, the instantaneous rate decreases with time, and consequently with $m_{Fe^{+++}}$. This relation between average and instantaneous rate is explained if it is assumed that the rate is proportional to the ferric—ferrous ratio, or possibly to the fraction of ferric iron present, rather than a function of $m_{Fe^{+++}}$ alone.

The relation between percent ferric iron remaining and time, for any given experiment, is like the relation between the percent of a radioactive element remaining and time. However, unlike radioactive decay, the "half life" of ferric iron is not independent of the original amount present.

Because the amount of ferric iron reduced per unit time is constant at a given ratio of $m_{Fe^{+++}}/m_{Fe^{++}}$, it appears that the rate-controlling mechanism may be related to adsorption of ferric and ferrous iron on the pyrite surface. If adsorption is rapid, and if the subsequent oxidation reaction is slow, then the reaction rate probably would be proportional to the fraction of the pyrite surface occupied by ferric ions (Fraction Fe^{+++} pyrite $= \frac{Fe^{+++}\text{ pyrite}}{Fe^{+++}\text{ pyrite} + Fe^{++}\text{ pyrite}}$).

As a first approximation, the fraction of the pyrite surface occupied by ferric ions might be expected to be roughly proportional to the fraction of ferric ions in the solution. If so,

$$\frac{dm_{Fe^{+++}}}{dt} = k_1 \frac{m_{Fe^{+++}}}{m^{\circ}_{Fe^{+++}}} = k_2 \frac{Fe^{+++}\text{ pyrite}}{Fe^{++}\text{ pyrite} + Fe^{+++}\text{ pyrite}}.$$

As a test of this relation, a number of runs are shown for Rio Tinto pyrite, in which $-\log m_{Fe^{+++}}/m^{\circ}_{Fe^{+++}}$ is plotted against time (fig. 9). All runs give straight line relations. The general equation that describes all the runs is

$$-\log \frac{m_{Fe^{+++}}}{m^{\circ}_{Fe^{+++}}} = \frac{k}{m^{\circ}_{Fe^{+++}}} \text{ time.}$$

Inasmuch as the ferrous and ferric ions have different charges, it might be expected that the adsorption of these ions on a solid surface would not be a simple function of their relations in solution. However, it is possible to obtain essentially the relations observed by using an ion exchange equation, if appropriate empirical constants are applied.

The rate of reduction of ferric ion is, therefore, evidently proportional to the fraction of pyrite surface upon which this ion is adsorbed, and to the total initial surface available. Apparently the fraction of surface occupied by ferric ion depends upon competition between ferric and ferrous ions, and is proportional to the fraction of ferric ion in solution. It seems likely that the fraction of ferric ion in solution, while proportional to that on the pyrite surface, is larger. This accounts for the marked increase in oxidation rate of pyrite at ferric fractions in solution very close to unity—the fraction occupied surface probably increases disproportionately as the ferrous concentration becomes vanishingly low.

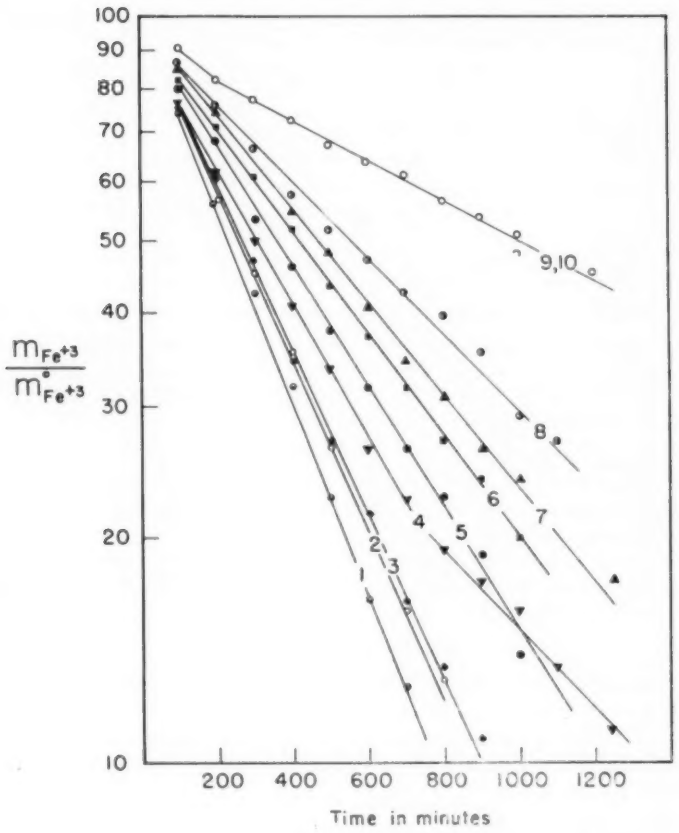

Fig. 9. Plot of log fraction ferric remaining $\frac{m_{Fe^{+++}}}{m^{\circ}_{Fe^{+++}}}$ versus time for some runs with Rio Tinto pyrite. (1), (2) $m^{\circ}_{Fe^{+++}} = 4 \times 10^{-4}$m/1; (3), (4), (5), (6), (7), (8), $m^{\circ}_{Fe^{+++}} = 1 \times 10^{-3}$m/1; (9), (10) $m^{\circ}_{Fe^{+++}} = 2 \times 10^{-3}$m/1.

Attempts to achieve equilibrium.—If equilibrium between pyrite and a ferrous-ferric solution could be achieved, it might be possible to deduce the oxidation reaction involved. But as shown in figure 7, which is typical of many similar experiments, no constant ferrous—ferric ratio is obtained. The Eh of the solution becomes unsteady in the vicinity of 0.250-0.300 volts. In part the explanation may be in the experimental conditions, for at an Eh of 0.250, in a solution originally 0.001m in iron, $m_{Fe^{+++}}$ is only about 10^{-7}. In such a solution, any minor impurities that might oxidize or reduce iron can change the measured potential many millivolts.

If the apparent equilibrium represented by the temporary asymptote in figure 7 is used as an equilibrium Eh, the reaction

$$FeS_2 = Fe^{++} + 2S^\circ_c + 2e \qquad (4)$$

describes the relation well. The observed Eh is about 0.250 volts, the calculated value the same.

The general picture of pyrite oxidation by ferric ion, as we now see it, is that the reaction is described through most of its course by oxidation to Fe^{++} and SO_4^{--}.

$$8H_2O + FeS_2 + 14Fe^{+++} \rightarrow 15Fe^{++} + 2SO_4^{--} + 16H^+ \qquad (5)$$

Oxidation to sulfate ion is irreversible. This mechanism is checked by allowing pyrite to be oxidized, and then re-titrating the solution with ceric sulfate. It is found that the iron content of the solution is increased by about the required 7 percent, showing that the sulfur has been carried up to sulfate. Only at very low potentials can reaction (4) hold.

Oxidation of pyrite by ferric ion near equilibrium thus seems to yield molecular sulfur. Any excess of ferric ions will oxidize this sulfur through intermediate species to sulfate ion. There may be a restricted Eh range in which the temporary oxidation product is an intermediate sulfur species, but it appears that under most natural conditions ferrous ion and sulfate ion are produced essentially instantaneously. This corroborates the findings of Temple and Koehler (1954, p. 24).

SUMMARY AND CONCLUSION

Oxidation of pyrite specimens by ferric sulfate solutions shows that pyrites from different localities oxidize at markedly different rates, but apparently the mechanism is the same. The rate of oxidation is chiefly a function of the oxidation potential of the solution, and is independent of the total iron content.

The oxidation process, over the range of ferric ion concentration from 100% to less than 0.1 percent $m_{Fe^{+++}}$, apparently is controlled by differential adsorption of ferric and ferrous ions on the pyrite surface. All our data are explained if it is assumed that the rate of oxidation is proportional to the fraction of pyrite surface occupied by ferric ion, and if the occupation of pyrite surface by ferric and ferrous ions is approximately proportional to their concentrations in solution. Thus if pyrite is placed in a solution containing ferric and ferrous ions, rapid adsorption of both species apparently takes place on the pyrite surface. Oxidation to ferrous ions and sulfate ions occurs only at those sites occupied by ferric ions. Because the oxidation process is slow relative to adsorption, the adsorption process controls the rate.

Impurities in solution such as nickel, cobalt, vanadium, manganese, copper, and cerium have little effect on oxidation rate, with just a hint that vanadium accelerates oxidation and copper retards it. The oxidation rate is independent of pH in the range 0 to 2. (No ferric hydroxides are formed.)

Except under near-equilibrium conditions, pyrite is apparently oxidized by ferric ion to ferrous ion, hydrogen ion, and sulfate ion. The first step of the oxidation process, however, may be to produce molecular sulfur.

Because of the exponential increase in oxidation rate as the ratio of ferric ion to total iron increases, pyrite changes from an essentially inert substance at Eh values in the vicinity of 0.300-0.400 volts, to an active reducing agent at potentials of 0.800 volts and above.

ACKNOWLEDGMENTS

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A FOLDED THRUST IN NEVADA—INFERENCES AS TO TIME RELATIONS BETWEEN FOLDING AND FAULTING*

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ABSTRACT. A segment of the Roberts thrust, a fault of regional extent in central Nevada, is strongly overturned at the border of a window in the northern Shoshone Range. The several branch faults that spring from the main thrust at and near the axis of overturning are, in order of decreasing age, successively less warped. This sequence suggests that folding of the thrust was concurrent with its development. The map pattern resembles that of many other windows in large thrusts: it is suggested that "eyelid" windows (which are framed by two or more thrust surfaces) are generally formed by concurrent folding and thrusting.

INTRODUCTION

The Roberts thrust of central Nevada has now been recognized over an area more than 100 miles along the strike and 50 miles broad (Roberts, Hotz, Gilluly and Ferguson). Its age appears to be Early Mississippian, though it has perhaps been rejuvenated in post-Triassic time. The most westerly windows thus far recognized in the thrust are on the west flank of the northern Shoshone Range, Nevada in the Mt. Lewis quadrangle (fig. 1).

The most northerly of these western windows, the Goat Ridge window, exposes some unusual fault relations which furnish the subject of this paper. The interest here lies solely in the light that the geometry of the faults may shed on the time relations between folding and faulting; the detailed stratigraphy and structure of the area are described elsewhere (Gilluly and Gates, in preparation).

THE GOAT RIDGE WINDOW

The general relations of the Goat Ridge window are shown in figure 2.

The "eastern facies" of Paleozoic rocks, which makes up the lower plate of the thrust, includes rocks of Cambrian, Ordovician and Silurian ages. These rocks closely resemble strata of equivalent ages in the standard Eureka section (Nolan, Merriam and Williams) and farther east in the Great Basin in being composed predominantly of carbonate rocks. The only notable noncarbonate rocks are the widespread Eureka quartzite of Middle Ordovician age and the local phyllites and greenstones of the Middle Cambrian.

The rocks of the "western facies", which make up the upper plate of the thrust and thus must have moved relatively eastward over the carbonate facies, contrast dramatically with those of the eastern facies. They are almost wholly siliceous: quartzite, chert, greenstone, sandstone, with subordinate shale and siltstone. Locally only Ordovician and Devonian rocks are represented in the upper plate, although elsewhere two distinct formations of Silurian sandstone and shale have also been recognized in it. The facies contrast is so clear that nowhere is there any question as to which fault plate is represented.

The Goat Ridge window also contains two small stocks of porphyritic granodiorite and many dikes of quartz latite porphyry. Only the stocks are shown in figure 2. These rocks are presumably of Tertiary age; their only note-

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Fig. 1. Index map showing location of northern Shoshone Range, Nevada.

worthy feature in connection with this paper is that, although most of the dikes trend generally north, the east-trending Trout Creek tear fault is also occupied (or followed) for several miles by them. The intrusions are all much younger than the faulting, as these and other masses nearby cut all the thrusts without offset.

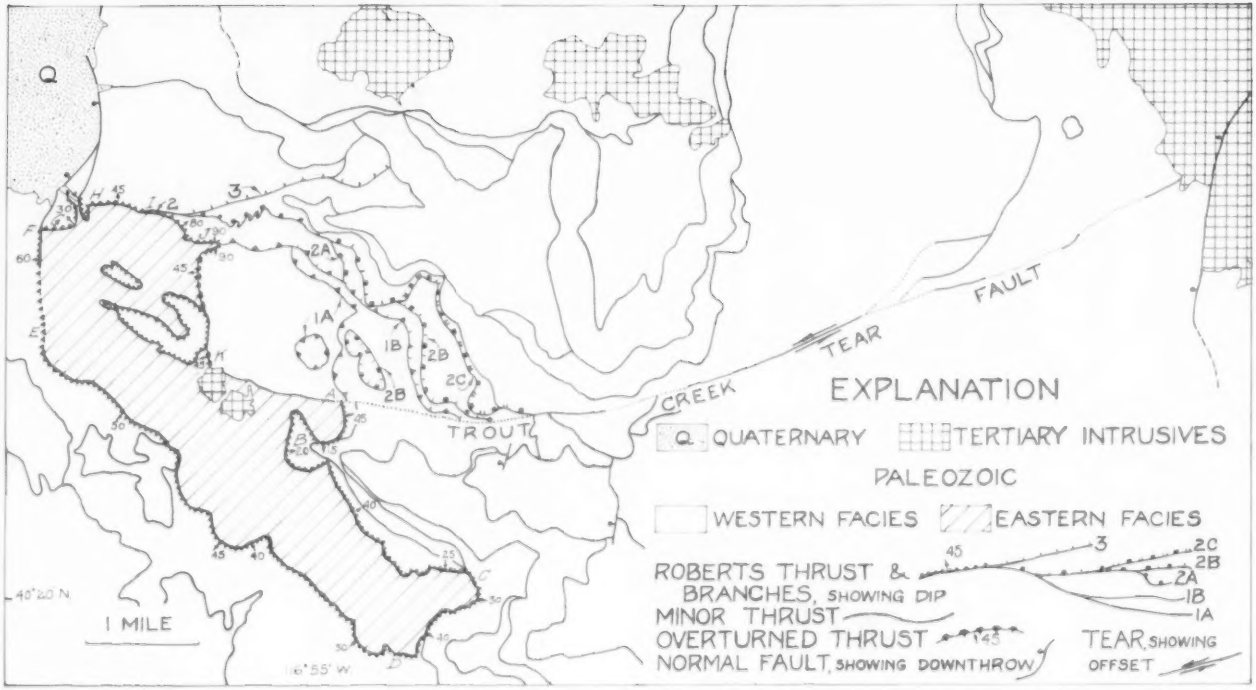


Fig. 2. Generalized geologic map of the Goat Ridge window in the Roberts thrust, Mt. Lewis quadrangle. The patterns on the branch faults are keyed to the inferred sequence of their origin.

The geometry of the window boundary is here summarized, beginning at the point marked A in figure 2, where the Roberts thrust abuts the south side of the Trout Creek fault, and thence proceeding in a clockwise direction. At A, which is just north of the crest of Goat Ridge, the thrust dips eastward at about 45 degrees. As the fault is followed to B, on the ridge crest, the dips become more gentle and the fault is seen to be curved in a shallow "synform" plunging gently eastward. From B to C, as the fault is followed down the south side of Goat Ridge, the dip is 40 to 25 degrees northeast and north. At C, where the fault crosses Mill Creek North Fork, it curves fairly sharply in an "antiform" that plunges eastward. Thence, past D, where its trace turns northwest to the mountain front at E, the fault dips southerly, chiefly to the southwest, at angles of 40 to 50 degrees. The north-south segment of the fault from E to F dips about 60 degrees west; perhaps for part of this stretch it coincides with a branch of a Basin Range fault, of Tertiary age. At G, at the northwest corner in the fault trace in the bottom of Trout Creek Canyon, the fault dips northwest at about 30 degrees. From this point eastward, past H along the north wall of

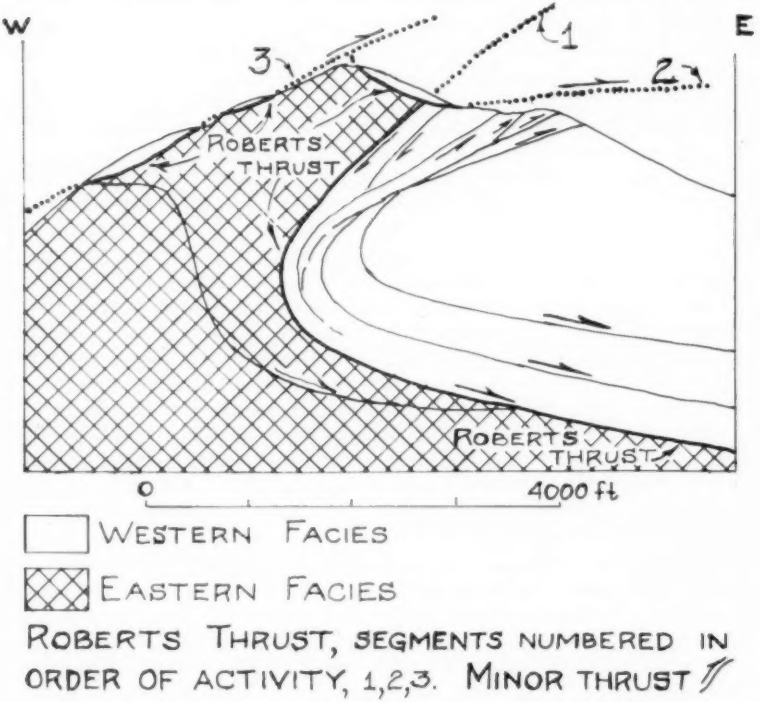


Fig. 3. Generalized east-west section through the north side of Goat Ridge about 2000 feet north of the crest, showing relations of various segments of the Roberts thrust. (The details shown here do not rigorously agree with those shown on fig. 2 because of the necessity for generalization in that figure.)

Trout Creek Canyon, the fault dips north and steepens. Within less than a mile, at I, it has become almost vertical and retains this attitude to the sharp hairpin curve at J, that terminates this segment of the trace. The hairpin marks a complete 180-degree turn in the fault trace, but the fault remains vertical and can be readily followed, with this attitude, across Trout Creek for about a quarter of a mile back westward. The trace thence can be followed almost directly south, up the steep south wall of Trout Creek to K, at the summit of the ridge, a distance of more than a mile horizontally, and nearly 3,500 feet vertically. A part of the trace near the hilltop is covered by a klippe of western facies whose relations are shown in figure 3. Clearly this mass moved into position after the fault was over-turned. The fault projection beneath this klippe can be validated because, both to the south near the porphyry stock, and to the north, as far as the sharp curve at the valley bottom near J, the Roberts fault dips westward at 40 to 50 degrees in very clear cliff exposures. In other words, the fault between J and K is overturned so that the rocks of the "window" have come to overlie those of the "frame."

At the south end of this segment, near K, the fault trace is cut by the western of the two stocks of granodiorite porphyry. Exposures here are excellent and the fault does not continue on trend southward beyond the stock. But, leading eastward out of the same mass, the contact between window and frame is the western end of the Trout Creek fault, which stands nearly vertically, trends east, and is occupied for about three miles by a series of quartz latite dikes. No extension of the Trout Creek fault beyond the stock can exist to the west; exposures are excellent and prove that it does not cut the rocks of the lower plate west of the porphyry mass. To the east, the Trout Creek fault extends in a gentle arc for about ten miles, to the quartz diorite stock of Granite Mountain, where it probably originally joined a large thrust in the upper plate of the Roberts fault. For about half this distance it is occupied by dikes.

This vertical fault—the Trout Creek—thus bounds the window on the north for about a mile and a quarter (from K to A in fig. 2); to the south, the thrust at its contact at A dips eastward as noted above, whereas to the north the Roberts fault is overturned and dips westward to 40 to 50 degrees.

This offset and the reversal of dip of the thrust (at K as compared with that at A, fig. 2), of course suggests that the Trout Creek fault is a tear fault, with left lateral movement that must increase downward. This inference is fortified by the difference in attitude of the minor (some nevertheless measurable in miles) thrusts in the upper plate of the Roberts thrust to the north and south of the Trout Creek fault, shown as simple lines in figure 2. To the south, in a belt several miles wide along the fault, these subsidiary faults in the upper plate dip at generally moderate angles of 20 to 45 degrees, they are only gently folded; to the north, several such thrust scales have been folded into isoclinal north-trending folds. These isoclinally folded thrust plates plunge northward at angles as steep as 50 degrees. The shortening exhibited by rocks of the upper plate north of the Trout Creek fault is thus several times that of comparable tectonic horizons for several miles south of the fault.

That the overturn in the Roberts thrust to the north of the Trout Creek fault is directly related to left-lateral movement along this fault seems to me

to be implied by the shape of the thrust fault and by the pattern of the branches that spring from it at and just west of the hairpin turn in its trace, at J.

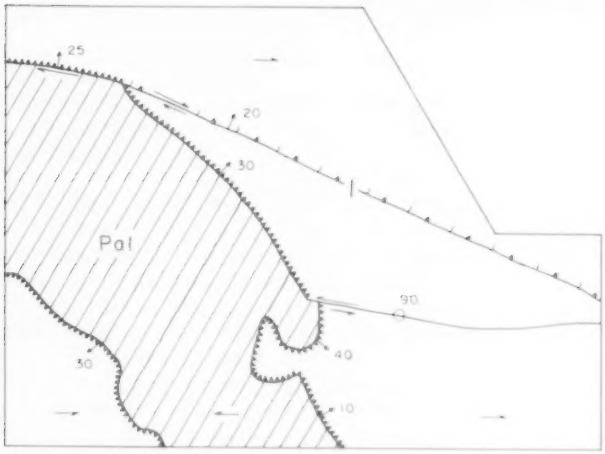
PATTERN OF BRANCH FAULTS

From the north side of the hairpin curve of the Roberts thrust (J in fig. 2) a branch fault springs into the upper plate along an easterly course. Half a mile from the trace of the Roberts thrust this fault in turn branches, one segment trending southeast and then south to a junction with the Trout Creek fault (1A on fig. 2), the other branch (1B on fig. 2) continuing the easterly course for about a thousand feet before also swinging southeastward to the Trout Creek fault. A short distance west of the hairpin curve, a second major branch (2 in fig. 2 and in fig. 5E) leaves the Roberts thrust. It shows notable drag folding, doubtless induced by the overriding of Branch 3 (fig. 2), just to the north, but trends generally eastward. About a mile to the east, a split of this fault (2A in fig. 2) curves sharply south to join the Branch 1B mentioned above. The other split curves tortuously southward to a junction with the Trout Creek fault, but splits off Branch 2C in its hanging wall a mile from its end.

At a point a few hundred feet west of the second split from the Roberts thrust, still a third, (Branch 3, figs. 2 and 5E) considerably straighter branch, leaves the main thrust and extends east-northeast for about three miles, to a point where it is cut off by a higher subsidiary thrust in the upper plate. The rocks overlying this branch for about two miles to the east and the main Roberts thrust to the west as far as the mountain front—a total distance of more than three miles—are of Devonian age. All the other rocks of the upper plate in this area are of Ordovician age. The minimum displacement on this branch must thus be several miles.

The pattern of these faults seems to indicate the sequence and manner of their formation. As in most thrust zones, the beds involved at shallower levels evidently met less resistance to travel than those at greater depth. They seem to have acted as "sliding sledges" on the rocks beneath, dragging the adjacent underlying rocks with them as they travelled forward. The fault surface thereby became bent (perhaps concurrent compression of the whole mass operated to emphasize the bending) along with the rocks above and below. Doubtless the main slip remained localized on the original shear surface while bending remained small but, after the bend became sharp enough, a new shear developed on a straighter course. In this way successive branches formed and became folded as new ones in turn formed at structurally higher levels while the faulting continued. I believe the fault pattern shown in figures 2, 4 and 5 is a strong argument for the sequence in development indicated by the numbers in these figures and also for the inference that their folding was concurrent with the folding of the thrust.

All of this sequence seems consistent with the left lateral sense of motion of the Trout Creek fault. If it be considered that the upper plate encountered more resistance to its eastward travel north of the line of the present Trout Creek fault than to the south of that line—a difference that may be reflected in the steep northward plunge of all the isoclinally folded thrust scales in that di-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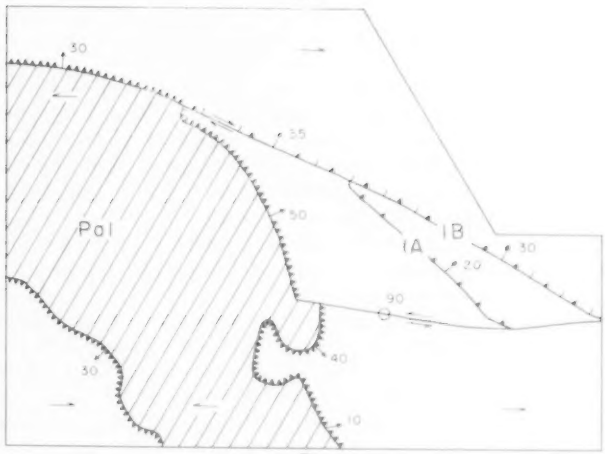
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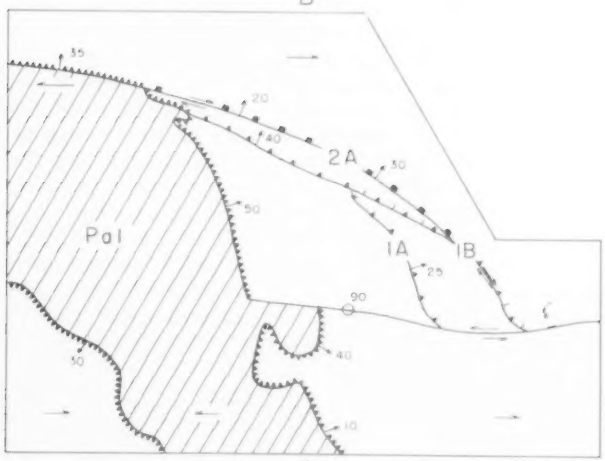
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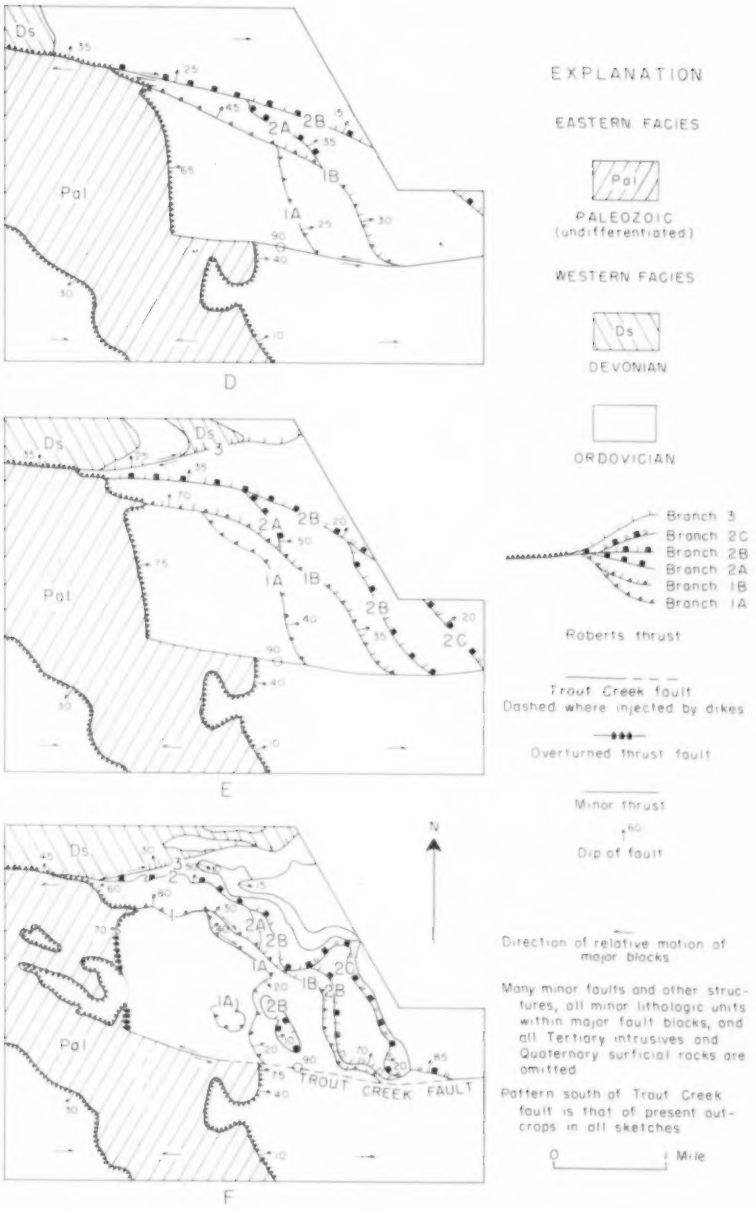

Figure 5 - Later stages in development of Goat Ridge overturn

Fig. 4. Earlier stages in development of Goat Ridge overturn.

rection—one can imagine successively higher diversion of the active slide planes as the rocks both below and above the thrust, and the thrust itself, became more and more tightly folded. Such a sequence is the one imagined and portrayed in figures 4 and 5.

All of the structures illustrated seem to have formed at depth. There is no sign, here or elsewhere so far as now recognized, of any surficial deposits being overridden by the advancing thrust sheets. Indeed, the highly plastic behavior of the thick quartzites, greenstones and sandstones involved implies a considerable depth of burial during the deformation.

EXTRAPOLATION TO OTHER WINDOWS

North of the Trout Creek fault, both the Roberts thrust and its branches and the higher scales in the upper plate of the thrust dip to the north. This direction of plunge abruptly stops at the Trout Creek fault; to the south the folds are gentle and irregular, some south and some north at low angles. It seems clear that all these features are mutually related: the folding of the Roberts thrust and its branches, the movement along the Trout Creek fault and the northward plunge of the structures to the north of the Trout Creek fault must all have developed together; they could hardly have developed in this close association if they were not genetically related. The steep northward plunge, by placing all the elements close together, permits analysis of their interconnection.

If it be granted that the folding and thrusting of the rocks of the Goat Ridge window probably occurred simultaneously, comparable map patterns around other windows may also indicate simultaneity.

An example of a similar pattern which has been similarly interpreted has been mapped and described by Ferguson and Muller (1949, plate 3) from the Pilot Mountains of Nevada. The Mac and Spearmint faults are warped and folded much like the Roberts thrust and its branches at Goat Ridge.

Another example is that of the Hot Springs window, in Tennessee and North Carolina, described by Oriel (1951) and shown in figure 6. This window is bordered by four faults and a still older one lies wholly within it. Such windows, framed by more than one thrust, have been called "eyelid" windows by Oriel. It seems clear that this arrangement of faults is analogous to the pattern of thrust and branches in the Goat Ridge window and can be interpreted as formed by continued folding of a single thrust surface by drag of an overriding mass. After the thrust became too sharply curved longer to serve as a gliding surface, a higher break-through took place near the line of inflection. Similarities to the classic Engadine window of the central Alps have been pointed out by Oriel (p. 24-26).

The Hot Springs window seems to be representative of a good many others in the southern Appalachians: the Chattanooga and associated faults in the Kingston quadrangle, (Rodgers, 1952, pl. 7); the Unaka Mt. and Limestone Cove and the Iron Mt.-Little Pond Mt. faults, all in the Roan Mt. quadrangle (Rodgers, 1952, pl. 5); the Dunham Ridge fault and its split near Chucky, Greenville quadrangle (Rodgers, 1952, pl. 4) and the Dunplin Valley and

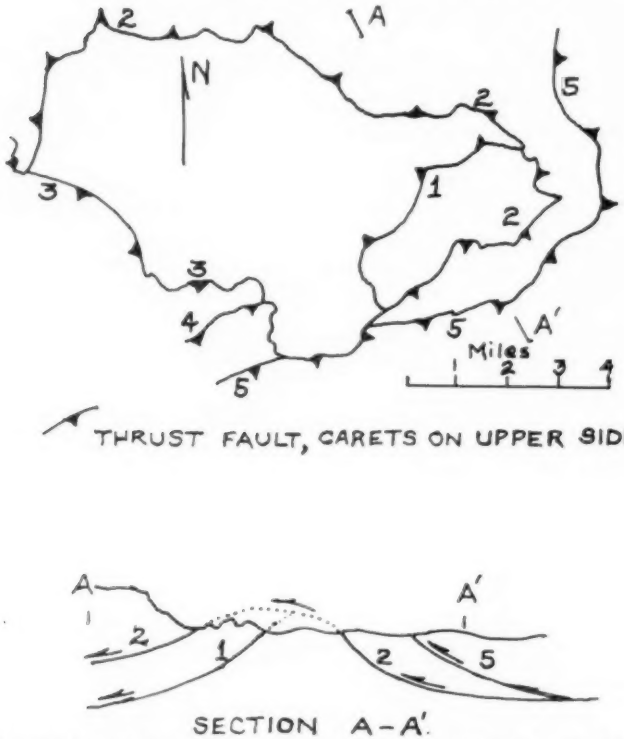

Fig. 6. Generalized sketch map of the Hot Springs window, after Oriel. The faults are numbered in order, from oldest to youngest.

Saltville faults in the Maynardsville quadrangle (Rodgers, 1952, pl. 2), all in Tennessee. The splits in the Cherokee block and overriding of the Buffalo Mountain block in northeastern Tennessee (Ordway, 1959) seems also to record concurrent folding and faulting.

Similar patterns, some but not all of which have been read in the way here suggested, have been described for the Condroz thrust of Belgium (Fourmarier, 1954, p. 674-677), for several faults in the Ruhr coalfield (Kienow, 1955, fig. 9, p. 148; Oberst-Brink, 1938), and for several faults of the Northwest Highlands of Scotland (Peach and others, figs. 10, 11, 12, 17, 18, 19). No doubt many others can be recognized elsewhere.

That such structures are, like those of Goat Ridge, formed by concurrent folding and thrusting is not, of course, proven by these analogies. Nevertheless the likelihood of their formation in this manner seems to me great enough to make improbable the conclusion of Keith (1928, p. 330) that the folding of the Appalachian overthrusts was all subsequent to their activity and indeed due to a later compressional episode. It seems much more likely that the warp-

ing of most of the folded thrusts took place concurrently with their activity as surfaces of transport.

The contrast in patterns to be expected in the two cases are schematically represented in figure 7. In the one, the plates of a sucession of thrusts all show

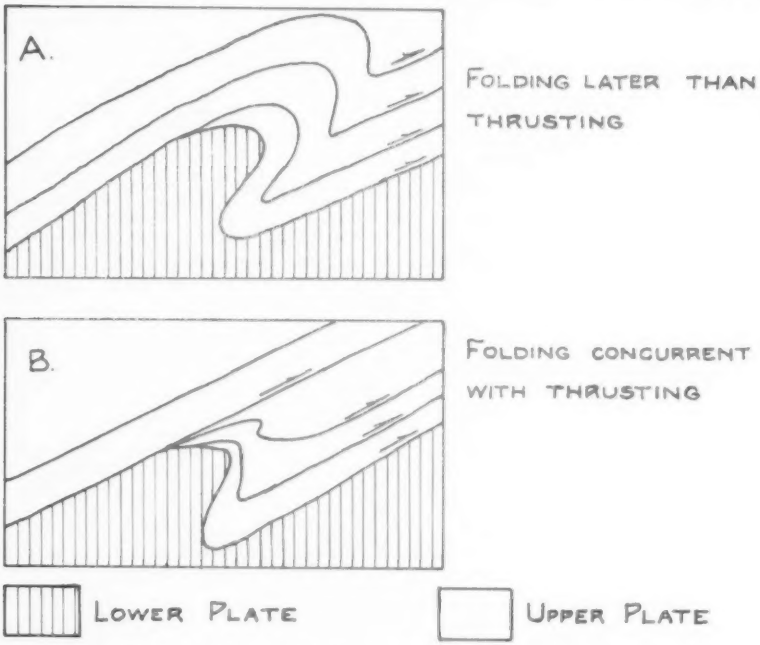

Fig. 7. Patterns to be expected if: A. Folding of thrust sheets is later than the thrusting; B. folding and thrusting are concurrent. The pattern can be considered as either a section of a structure whose tectonic axis is horizontal, or a map (and section) of a plunging structure.

essentially parallel convolutions; in the other, successively higher fault plates show simpler and simpler folds. That the second case exists seems to me probable from the relations here described on Goat Ridge in the Shoshone Range. The first case has been well described and mapped by Bertrand (1900) from the Gard coalfield of the Cevennes, and other excellent examples are known from the western Alps (Gignoux and Moret, fig. 20 and p. 223), and Alberta (Hume, 1941).

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INTERPRETATION OF EROSIONAL TOPOGRAPHY IN HUMID TEMPERATE REGIONS*

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ABSTRACT. Since the period 1890 to 1900 the theory of the geographic cycle of erosion has dominated the science of geomorphology and strongly influenced the theoretical skeleton of geology as a whole. Some of the principal assumptions in the theory are unrealistic. The concepts of the graded stream and of lateral planation, although based on reality, are misapplied in an evolutionary development, and it is unlikely that a landscape could evolve as indicated by the theory of the geographic cycle.

The concept of dynamic equilibrium provides a more reasonable basis for the interpretation of topographic forms in an erosionally graded landscape. According to this concept every slope and every channel in an erosional system is adjusted to every other. When the topography is in equilibrium and erosional energy remains the same all elements of the topography are downwasting at the same rate. Differences in relief and form may be explained in terms of spatial relations rather than in terms of an evolutionary development through time. It is recognized however that erosional energy changes in space as well as time, and that topographic forms evolve as energy changes.

Large areas of erosionally graded topography in humid regions have been considered to be "maturely dissected peneplains." According to the equilibrium theory, this topography is what we should expect as the result of long continued erosion. Its explanation does not necessarily involve changes in base level. Pediments in humid regions and some terraces are also equilibrium forms and commonly occur on a lowland area at the border of an adjacent highland.

INTRODUCTION

The part of geologic theory that deals with the interpretation of landforms and the history of landscape development has been dominated for several generations by the ideas of William Morris Davis and his followers. Davis' theory of landscape evolution was first fully presented in his essay, "The Rivers and Valleys of Pennsylvania" (Davis, 1889).[1] The important concepts that he introduced include the geographic cycle, the peneplain, and the formation of mountains by a succession of interrupted erosion cycles. Davis' theories became immensely popular among geologists in Europe as well as in America, though there were dissenters, including, for example, Tarr (1898) and Shaler (1899). His theory of the evolution of mountains as topographic features through the mechanism of multiple erosion cycles was especially influential and came to have a great influence on the theoretical skeleton of the whole science of geology. Its impact is still felt. Many of our ideas relating to the history of mountains, the internal constitution of the earth and the origin of some ore deposits are closely related to this theory. The idea that mountain ranges are vertically uplifted after they have been folded was conceived in order to explain the widespread existence of dissected peneplains (Daly, 1926). Another example is the theory of origin of bauxite and of manganese ores and other residual concentrates in the Appalachian Highlands, that are thought by some to have formed on a Tertiary peneplain surface (Hewett, 1916; Stose and Miser, 1922, p. 52-55; Bridge, 1950, p. 196.)

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[1] Davis' major papers dealing with the sculpture of landscapes by streams (Davis, 1889, 1890, 1896a, 1896b, 1899a, 1899b, 1902a, 1902b, 1903, 1905a, 1905b) as well as others of his papers were collected in one volume published as "Geographical Essays" (Davis, 1909) and reprinted in 1954 (Davis, 1954). The 1954 edition of "Geographical Essays" has the same page numbers as the 1909 edition.

In the last 20 years, however, Davis' ideas have become less popular and the small but ever-present number of geologists who were skeptical of his theories has increased. Though many geologists have been dissatisfied with it, the theory of the geographic cycle and its application to the study of landforms has not generally been replaced by any other concept. Several alternative theories have been proposed, including the theory of Penck (1924, 1953) which relates the form of slopes to changes in the rate of uplift relative to the rate of erosion, and the "pediplain" theory of L. C. King (1953), an elaboration and expansion of Penck's concept of slope retreat. Both of these theories, however, are also cyclic concepts and hold that the landscape develops in stages that are closely dependent on the rate of change of position of baselevel.

During the course of my work in the Central Appalachians which began in 1952, seeking a different approach to geomorphic problems, a conscious effort was made to abandon the cyclic theory as an explanation for landforms. Instead, the assumption was made that the landforms observed and mapped in the region could be explained on the basis of processes that are acting today through the study of the relations between phenomena as they are distributed in space. The concept of dynamic equilibrium forms a philosophical basis for this kind of analysis. The landscape and the processes molding it are considered a part of an open system in a steady state of balance in which every slope and every form is adjusted to every other. Changes in topographic form take place as equilibrium conditions change, but it is not necessary to assume that the kind of evolutionary changes envisaged by Davis ever occur. The consequences and results of this kind of analysis in most cases differ from conclusions arrived at through the use of the cyclic concepts of Davis.

On rereading some of the classic American literature in geomorphology I realized that G. K. Gilbert used essentially this approach and that I have followed a way of thinking inherited either directly from him or from some of his colleagues. Even though Davis and Gilbert were contemporaries and friends, Gilbert makes little use of and few references to the theory of the geographic cycle or any of its collateral ideas. This omission is so conspicuous that it is difficult to believe Gilbert ever wholeheartedly accepted the idea. It seems to me that Gilbert's famous paper, "Geology of the Henry Mountains" (Gilbert, 1877, p. 99-150) outlines a wholly satisfactory basis for the study of landscape that does not foreshadow the developments in geomorphology that followed in the next 50 years.

In the pages that follow some concepts inherent in the theory of the geographic cycle that seem to me unsound are briefly discussed. The alternative approach to landscape studies based on spatial relations in a system in equilibrium is briefly presented. Very few of the ideas are original and most of them have been published in one form or another in the works of other geologists. In addition I wish to acknowledge the considerable assistance obtained in stimulating discussions with my friends and colleagues, especially C. S. Denny, J. C. Goodlett, C. B. Hunt, L. B. Leopold, C. C. Nikiforoff, and M. G. Wolman with whom I have been associated at various times during the formulation of these ideas. The manuscript has been read and criticized by R. P. Sharp of the California Institute of Technology, Sheldon Judson of Princeton University,

C. C. Nikiforoff (formerly of the Department of Agriculture) as well as by some generous colleagues in the U. S. Geological Survey.

THE GEOGRAPHIC CYCLE AND THE PENEPLAIN CONCEPT

The theory of the geographic cycle rests on the assumption that there is a base level toward which every area erodes and to which the streams become graded. After an initial uplift or rise of a part of the earth's crust, erosion proceeds through successive stages of youth, maturity, and old age; from a time in which stream grades are irregular, through a time when they become smooth, to a stage of low relief when the entire landscape is reduced close to base level. An important stage in the cycle is reached when the slopes of the larger streams are so reduced that they are able to transport just the amount of debris supplied from upstream and no more. At this point the stream is said to be graded and the stage of maturity begins. The trunk streams, unable to erode their beds, shift laterally, forming floodplains and meanders. The debris-covered area on the valley floor expands laterally as the interstream divides are lowered. At the final stage of old age the landscape is one in which the streams meander across broad plains covered by a sheet of waste. The divide areas are graded to the streams and are covered by a waste mantle transported by creep. They rise only slightly above the shallow valleys (Davis 1909, p. 254-272; 1899, p. 485-499). Davis envisaged that there must be interruptions of the ideal cycle of erosion and in fact that an ideal cycle is rarely completed. Alternate periods of uplift and stability of base level result in successive incomplete cycles during which the uplifted peneplains are dissected and new ones form along the streams. Hilly areas in which the hilltops rise to roughly the same height above the streams, like the Piedmont region of the central Atlantic States, are peneplains that have been uplifted and dissected by stream erosion until a stage of maturity has been reached (Davis, 1909, p. 272-274; 1899, p. 499-501).

The concept of planation.—The concept of *planation* or *lateral planation* is lucidly presented in the Henry Mountain report (Gilbert, 1877, p. 127). Gilbert described the process of planation in connection with the formation of smoothly graded, gravel-covered surfaces, cut on soft Mesozoic rocks at the foot of the Henry Mountains. Surfaces like these have come to be known as *pediments* and have been restudied in the Henry Mountains by Hunt, Averitt, and Miller (1953). It was recognized by Gilbert that the planation in this region occurs on soft rocks, such as weak sandstones and shales where slopes or declivities are small in comparison with the trachyte mountains in which the streams originate. Gilbert thought that lateral shifting of the streams is dependent on the fact that the bed load transported by the stream is more resistant than the rock through which it flows so that the stream cuts laterally against the soft bank. Where one of these streams cuts again through hard rock lateral planation ceases (Gilbert, 1877, p. 130) and canyons are formed. The process is dependent on the geology of the drainage basin of the laterally shifting stream, and on a contrast in rock resistance and slope between the upper and lower parts of the basin.

Davis applied the erosion cycle concept to the idea of lateral planation. In his theory a stream always has a tendency to erode laterally against its banks.

As the landscape passes through the evolutionary stages of the erosion cycle, first the larger streams and later the tributaries approach the base level of erosion. As they do so their ability to cut downward diminishes. They migrate laterally eroding the valley walls, producing a floodplain or surface of planation.

It is interesting to note the contrast between the planation observed and described by Gilbert and the planation envisaged by Davis. Gilbert's explanation of lateral planation involves a dynamic equilibrium of forces existing at the present time in actual drainage basins and the relation of these forces to the rocks. Davis' theory on the other hand assumes that lateral planation occurs in any drainage basin with the passage of time, regardless of its geology. In fitting the concept of planation into the framework of the geographical cycle Davis attempted to rationalize relations between things that change through time and hence cannot be observed or measured. In the transfer from a scheme of ideas that involves space to a scheme that involves time Davis ignored the spatial relations cited by Gilbert that make the concept valid.

Surfaces of planation are produced by streams under certain circumstances, but there is no reason to believe that such surfaces enlarge through time as relief is lowered, merely as a consequence of a reduction in slope. On the contrary it is likely that as gradation proceeds, the efficiency of the stream system in removing the waste of its drainage basin may increase.

The graded stream.—One of the key ideas in the theory of the geographic cycle is the concept of the graded stream. The word "grade" was used by Gilbert (1877, p. 112) in discussing the stable slope of the stream channel in the same sense that an engineer uses it to describe the slope of a railroad or highway. Davis borrowed the term at Gilbert's suggestion and used it in a more special sense to designate a certain stage in the evolution of stream profiles when the stream's ability to transport the load supplied to it from above is just balanced by the load that it has to carry (Davis, 1909, p. 392; 1902, p. 89).

This concept of grade has probably received more discussion among geomorphologists than any other aspect of the geographic cycle (for example Kesseli, 1941, Mackin, 1948, Woodford, 1951, Rubey, 1952, Leopold and Maddock, 1953, Wolman, 1955). As suggested by Kesseli, the concept as outlined by Davis seems rather elusive, so that it is difficult to identify a graded stream in nature. Mackin, however, in his study of the graded stream, clarifies some of the ideas and suggestions of Davis. The examples of graded streams cited by him are migrating laterally, depositing on the floodplain an amount of material equal to what they erode by lateral cutting. The graded stream is not actively cutting vertically downward and its longitudinal profile is being changed only very slowly as the relief or other conditions in the drainage basin change. Since it is cutting laterally, the channel of the graded stream is bordered by a floodplain underlain by thin river deposits and by terraces whose composition is entirely material carried from upstream and different from the underlying rock (Mackin, 1948, p. 472).

Leopold and Maddock (1953) considered the graded stream in relation to the hydraulic geometry of the channel. Their study of stream gaging and cross section data indicates just as consistent a pattern in the relationships between

the variables width, depth, velocity, and sediment load in ungraded as in graded streams. They conclude that Mackin's concept of grade cannot be demonstrated by consideration of stream gaging data and they use the term "quasi-equilibrium" in reference to the equilibrium in stream channels observed by them in all the streams studied. They recognize that this equilibrium is distinct from the equilibrium implied by Davis and Mackin in the concept of the graded stream.

In Davis' concept of grade, high velocity of flow and a high capacity are associated with a steep channel slope. As slope diminishes during the evolution of the landscape through the erosion cycle, velocity diminishes as well as the capacity to transport debris (Davis, 1909, p. 397-398; 1902, p. 95-96). This idea may have seemed reasonable to Davis because he shared with many others the belief that mountain streams with steep channels have higher velocities of flow and therefore greater capacity than do large streams with lower slopes in lowland areas. This observation is not necessarily true. Actual measurements in many natural streams demonstrate that for equivalent frequencies of discharge average velocities tend to increase downstream rather than decrease (Leopold, 1953). Studies of some Appalachian streams, furthermore, indicate that the size of material a stream has on its bed and banks is not related directly to slope, but is related also to discharge and other variables in such a way that in many streams the competence (or size of material that a stream can transport) increases downstream as slope diminishes (Hack, 1958). These facts make it appear doubtful that streams reach a balanced condition through any evolutionary sequence involving a gradual reduction in slope. Probably the balance that exists in most streams is the quasi-equilibrium described by Leopold and Maddock (1953, p. 51). This is a balance among at least seven variables. It is so complex and there are so many alternative adjustments possible that equilibrium can be achieved under many conditions and is arrived at very quickly, almost immediately, in the development of a valley. The uniform, or regular concave-upward longitudinal profile that is characteristic of many streams and has been called "the profile of equilibrium" results not from the attainment of a certain stage in the evolution of a valley, but merely from the regular change downstream in some of the many variables involved in channel equilibrium. Most important of these is probably discharge that increases downstream as a consequence of a regular enlargement of the drainage area.

The streams cited by Mackin (1948) as examples of graded streams represent special cases that are exceptional rather than general. Like Gilbert's streams in the Henry Mountains, such streams head in hard rock areas of high slope and altitude. Their lower courses are in soft rocks and as a consequence have a low slope relative to the increased discharge. They migrate laterally and have a diminishing competence only because of the geologic pattern of the terrain they traverse. They represent a class of streams intermediate between those whose competence with respect to the load derived from upstream is increasing in a downstream direction or remains the same, and those whose competence decreases downstream so abruptly that they aggrade actively enough to build fans. They are no more in a state of equilibrium or disequilibrium

than a mountain torrent that is engaged in cutting a gorge. The torrent also has a bed load and lag deposit or floodplain along the bank composed of material too coarse for the stream to move in the ordinary flood, but in this case the lag deposit is locally derived by washing or sliding down the adjacent slope or by plucking from the bed.

The stage of old age and the maturely dissected peneplain.—In the concept of the geographic cycle the appearance of the land surface in the stage of old age is dependent on the process of lateral planation. The ideal surface is a plain partly graded by planation and covered by a veneer of waste. Divide areas with convex upward slopes exist, but are relatively smaller in area than in earlier stages. Such graded surfaces, as stated above, do not now exist in nature. The extensive plainlands of the earth are either depositional surfaces like alluvial plains, deltas, drift plains, and coastal plains, or if they are erosion surfaces in humid areas, they are hilly with rounded divides and steep-walled valleys that have generally come to be described as "maturely dissected peneplains." Exceptions are pediments and terraces, that in humid regions occupy relatively small areas. Excellent examples of maturely dissected landscapes in America are found in the Piedmont region of eastern North America, or in the Central United States where the so-called Ozark peneplain has been "dissected and uplifted." Large areas of the Canadian shield have been said to be a dissected peneplain whose drainage has been disrupted by glaciation. The great limestone valley of the Appalachians, similarly, has been said to be a dissected peneplain as is the plateau area to the west of the Appalachians. Thus land surfaces that are worn down to the stage of old age, as conceived by Davis, are virtually nonexistent; on the other hand former old age surfaces that have been dissected to Davis' stage of maturity are ubiquitous in the older terrains of the earth, especially in humid regions. Indeed this kind of topography is so universal it suggests that the end product or end surface toward which erosion proceeds resembles the "maturely dissected" surface rather than the "old age" surface or peneplain. Such an end surface is one whose forms are graded for the efficient removal of waste rather than one on which the waste products accumulate and stagnate.

THE PRINCIPLE OF DYNAMIC EQUILIBRIUM IN LANDSCAPE INTERPRETATION

An alternative approach to landscape interpretation is through the application of the principle of dynamic equilibrium to spatial relations within the drainage system. It is assumed that within a single erosional system all elements of the topography are mutually adjusted so that they are downwasting at the same rate. The forms and processes are in a steady state of balance and may be considered as time independent. Differences and characteristics of form are therefore explainable in terms of spatial relations in which geologic patterns are the primary consideration rather than in terms of a particular theoretical evolutionary development such as Davis envisaged.

The principle of dynamic equilibrium was applied to the study of landforms both by Gilbert (1877, p. 123) and by Davis (1909, p. 257-261, 389-400; 1899, p. 488-491; 1902, p. 86-98). Recently Strahler has outlined the principle in more modern terms as it might be applied to landscapes (Strahler,

1950, p. 676). The concept requires a state of balance between opposing forces such that they operate at equal rates and their effects cancel each other to produce a steady state, in which energy is continually entering and leaving the system. The opposing forces might be of various kinds. For example, an alluvial fan would be in dynamic equilibrium if the debris shed from the mountain behind it were deposited on the fan at exactly the same rate as it was removed by erosion from the surface of the fan itself. Similarly a slope would be in equilibrium if the material washed down the face and removed from its summit were exactly balanced by erosion at the foot.

In the erosion cycle concept of Davis, equilibrium is achieved in some part of the drainage system when there is a balance between the waste supplied to a stream from the headwaters and the ability of the stream to move it, or in other words, when the slope of the channel is reduced just enough so that the stream can transport the material from above with the available discharge. As argued on page 9 this kind of equilibrium probably is achieved in a stream almost immediately and is not related to a particular stage in its evolution. Davis' concept would imply that some parts of a drainage system would be in equilibrium whereas at the same time other parts would not, and that the condition of equilibrium is in time gradually extended from the downstream portion to the entire drainage system.

Rather than a concept of balance between the load of a stream and the ability of the stream to move it, it is more useful in the analysis of topographic forms to consider the equilibrium of a particular landscape to involve a balance between the processes of erosion and the resistance of the rocks as they are uplifted or tilted by diastrophism. This concept is similar to Penck's concept of exogenous and endogenous forces (1924, 1953). Suppose that an area is undergoing uplift at a constant rate. If the rate of uplift is relatively rapid, the relief must be high because a greater potential energy is required in order to provide enough erosional energy to balance the uplift. The topography is in a steady state and will remain unchanged in form as long as the rates of uplift and erosion are unchanged and as long as similar rocks are exposed at the surface. If the relative rates of erosion and uplift change, however, then the state of balance or equilibrium constant must change. The topography then undergoes an evolution from one form to another. Such an evolution might occur if diastrophic forces ceased to exert their influence, in which case the relief would gradually lower; it might occur if diastrophic forces became more active, in which case the relief would increase; or it might occur if rocks of different resistance became exposed to erosion. Nevertheless as long as diastrophic forces operate gradually enough so that a balance can be maintained by erosive processes, then the topography will remain in a state of balance even though it may be evolving from one form to another. If, however, sudden diastrophic movements occur, relict landforms may be preserved in the topography until a new steady state is achieved.

The area in which a given state of balance exists and that may be considered a single dynamic system may be conceived as very small or very large. In the Appalachian region, it may be that large areas are essentially in the same state of balance. In the West, however, in an active diastrophic belt, a

single dynamic system may constitute only a small area such as a single mountain range or a small part of a mountain range. Furthermore, because of sudden dislocations of the crust relict forms may be preserved in the landscape that reflect equilibrium conditions that no longer exist.

The crust of the earth is of course not isotropic and within a single erosional system, no matter how small, there is a considerable variation in the composition and structure of the crust. These variations are reflected by variations in the topography. Consider, for example, an area composed partly of quartzite and partly of shale. To comminute and transport quartzite at the same rate as shale, greater energy is required; and since the rates of removal of the two must be the same in order to preserve the balance of energy, greater relief and steeper slopes are required in the quartzite area. Similarly geometric forms differ on different rock types. An area that is underlain by mica schist or other igneous or metamorphic rock subject to rapid chemical decay, has more rounded divides than an area underlain by qaurtzite, if both are in equilibrium in the same dynamic system, for the schist is comminuted by weathering to silt and clay particles that are rapidly removed from hill tops on low slopes. On the other hand to remove quartzite from a divide at the same rate, steeper slopes and sharper ridges are required because the rock must be moved in the form of larger fragments.

The analysis of topography in terms of spatial or time-independent relations provides a workable basis for the interpretation of landscape. This kind of analysis is uniformitarian in its approach, for it attempts to explain landscapes in terms of processes and rates that are in existence today and therefore observable. It recognizes that processes and rates change both in space and time, and, by clarifying the relation between forms and processes, it provides a means by which the changes can be analyzed.

THE RELATION OF SOIL TO TOPOGRAPHY

Cyclic concepts of soil evolution have developed in a manner parallel to the erosion cycle concept of geomorphology. The idea of a cyclic evolution of soil through a stage of maturity to senility in which the profile becomes intensified and thickened through time is dependent on the concept of a topography that is stable, such as might exist on a peneplain or on a remnant of a dissected peneplain. Naturally enough, this idea lends support to the cyclic concept of landscape evolution.

An alternative theory of soil evolution based on dynamic equilibrium has been forcefully presented by C. C. Nikiforoff (1942, 1949, 1955, and 1959, p. 188) and parallels the concept of equilibrium in landscape evolution. As explained by Nikiforoff, nearly all soils achieve a state of dynamic equilibrium if they are exposed to the surface for a sufficient time. Factors in the equilibrium include climate, slope, rate of erosion, composition of the parent material, vegetation, and others. The horizons of the soil become diversified and owe their existence to an equilibrium among processes that tend to accumulate certain substances at certain depths, and those that tend to remove them. Take the clayey "B" horizon as an example:

The cyclic viewpoint holds that the clay in the "B" horizon accumulates

through leaching of the "A" horizon above it. The concentration of clay increases through time, though at a slower and slower rate until further accumulation is impossible. At this point the soil is mature and remains in this state until removed by erosion.

In terms of dynamic equilibrium, on the other hand, the amount of clay present in any horizon of the soil is the result of a balance at that place between the rate of clay accumulation and clay removal. These rates differ in different horizons and subhorizons and the balance between them determines the amount of clay present. Similar balance between rates of removal and of accumulation and the interactions between them determine the composition of all the horizons.

From the point of view of landscape interpretation this view of the soil has many advantages. It permits the lowering of the hilltop by erosion at a more or less constant rate, at the same time maintaining the equilibrium of the soil. In fact the soil profile is dependent on erosion as one of the factors in the equilibrium.

Compare a hilltop underlain by pure limestone or by clayey or silty limestone with a hilltop nearby underlain by cherty limestone containing massive beds of chert (fig. 1). The hilltop underlain by silty limestone does not ac-

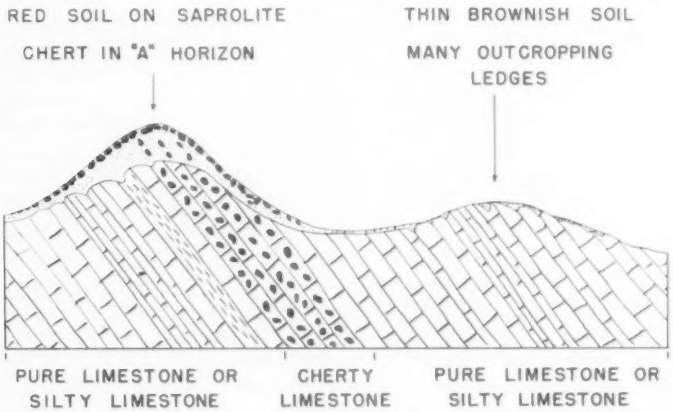

Fig. 1. Hypothetical cross section through two hills in a limestone region showing the relation of the bedrock and topography to the residual soils.

cumulate weathering products because they are removed by erosion as soon as they are freed from the rock by solution of the carbonate that binds them. On the other hill, underlain by cherty rocks, though solution goes on at the same rate, a debris of chert fragments is produced in addition to fine silt. These are too coarse to be removed by sheet erosion on slopes as gentle as those of the first hill. The chert accumulates forming a protective armor that prevents finer insoluble residues, also present, from being eroded. Accumulation of this deposit continues until the slopes are steep enough for the chert to be removed. The topography is now in equilibrium. One hill is covered with saprolite or

residuum and the other is not, but the saprolite-covered hill has steeper slopes and rises higher above the streams than the other (Hack, 1958b). It might be said that the residual material is "stored" on top of the hill as its covering armor is comminuted to sizes that can be removed by creep and wash. During the period of "storage" the material becomes oxidized and reddened and a profile develops. The time of storage may be very long, even thousands of years, and may be long enough for the soil to have survived major changes in climate and to owe some of its characteristics to irreversible reactions that took place in the past (Nikiforoff, 1955, p. 48).

EXAMPLES OF EROSIONALLY GRADED OR EQUILIBRIUM TOPOGRAPHY

As an area is graded by erosional processes the differences in the bedrock from one place to another cause a differentiation of the forms on them. Landscapes that develop on intricate and actively rising fault blocks may bear a closer relation to major structural features than to the underlying rock, but in a landscape like that of the Appalachian region in which large areas are mutually adjusted, the diversity of form is largely the result of differential erosion of rocks that yield to weathering in different ways. Such topography may be referred to as erosionally graded.

Ridge and ravine topography.—Many of the erosionally graded landscapes in humid temperate regions belong to the almost ubiquitous type that is commonly known as the "maturely dissected peneplain". Preferring a term that has no genetic connotation a more descriptive one such as *ridge and ravine* topography is suggested. This term refers to the monotonous network of branching valleys and intervening low ridges that make up the landscape of large areas. This topography may be concisely explained in terms of dynamic equilibrium in the words used by Gilbert (1909) in his discussion of rounded hilltops. He conceived that the important elements of the topography could be divided into two domains. The first, a domain of stream sculpture represented by channels in which the slopes are concave upward, because, as he says (Gilbert, 1909, p. 344), the transporting power of a stream per unit of volume increases with the volume; the transporting power also increases with the slope; and a stream automatically adjusts slope to volume in such a way as to equalize its work of transportation in different parts. The other domain is that of creep, represented by the slopes between the channels. In this domain slopes are mostly convex. Gilbert states (1909, p. 345) that:

> This is because the force impelling movement of material is gravity which depends for its effectiveness on slope. On a mature or adjusted profile the slope is everywhere just sufficient to produce the proper velocity. It is greatest where the velocity is greatest and therefore increases progressively with distance from the summit.

The forms of well-graded ridge and ravine landscapes vary within wide limits. Typical examples, both in areas of high relief, have been described by Strahler (1950) and by Hack and Goodlett (in press, 1960). Somewhat gentler topography of the same type is widespread in the Piedmont Province. An example of such an area is shown in figure 2. This is in a drainage basin tributary to the Patapsco River in Carroll County, Maryland. The bedrock is phyllite that is cut by veins of quartz. The interstream divides are convex upward

Fig. 2. Topographic map of area in the Piedmont Province of Maryland, showing typical ridge and ravine landscape (1950, U. S. Geol. Survey Winfield Quadrangle, Maryland).

and if measured on a coordinate system in which the origin, or zero point, is the top of the hill or ridge, they have the form of a parabolic curve like the one shown in fig. 3. They intersect the stream bottoms in steep slopes and

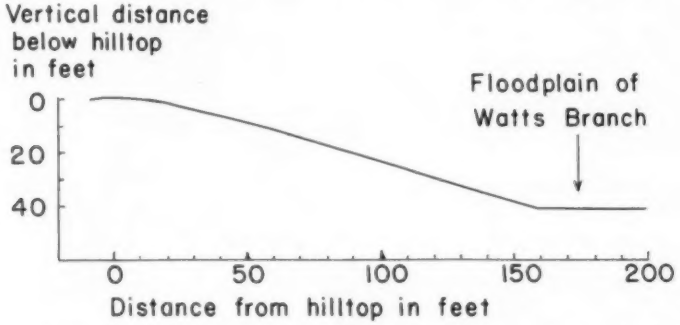

Fig. 3. Profile of hill slope near Rockville, Maryland, on fine-grained mica schist.

sharp angles, though in places the foot of the slope is concave upward, especially in slopes that intersect the floodplain of a stream at a point opposite the channel. The regularity of the landscape and the rather uniform height of the hills owe their origin to the regularity of the drainage pattern that has developed over long periods, by the erosion of rocks of uniform texture and structure.

Differences in form from one area to another, including the relief, form of the stream profiile, valley cross sections, width of floodplain, shape of hill tops and other form elements are explainable in terms of differences in the bedrock and the manner in which it breaks up into different components as it is handled on the slopes and in the streams.

Pediments.—Where differences in rock resistance in graded landscapes are slight or confined to narrow or small areas, the differences in topography are small. Where, however, two large areas, one of resistant rock and the other of much softer rock, are juxtaposed, the differences are not only pronounced but there is a zone of transitional forms on the less resistant rock. In both areas ridge and ravine landscapes are developed, but as the more resistant area has greater relief, steeper slopes, and sharper divides, debris is shed from the higher to the lower area. This kind of situation is a common one in the Appalachian Highlands where many valley or lowland areas are underlain by limestone and shale and are bordered by ridges or series of ridges underlain by sandstone and quartzite. The transitional forms are broadly fan-shaped gravel-covered and dissected surfaces cut by streams on bedrock that closely resemble typical pediments in many western areas. They are called pediments because of this resemblance, but it is recognized that similar surfaces may be produced by different processes.

An example of a pediment area in the headwaters of the South Fork, Shenandoah River, Augusta and Rockingham Counties, Va., may clarify the equilibrium relations involved. As shown in figure 4, Dry River, Briery Branch,

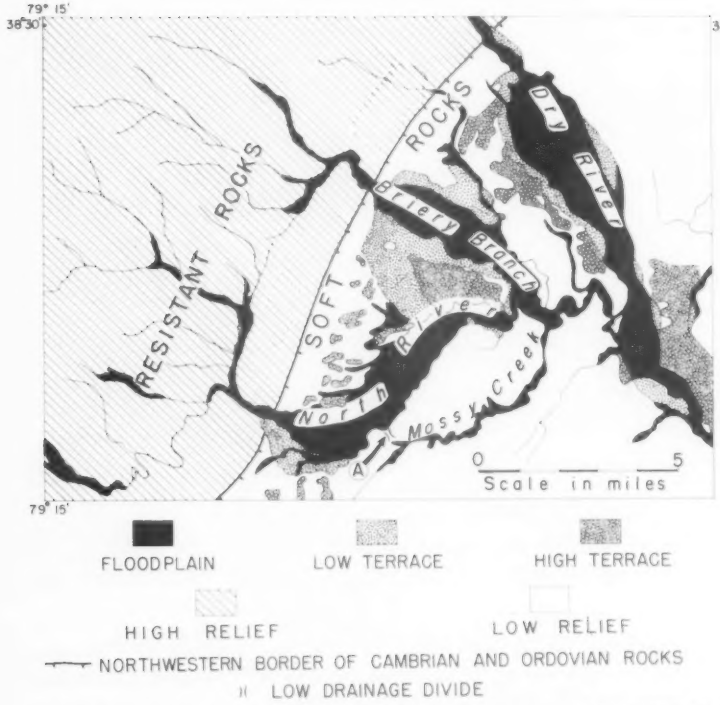

Fig. 4. Simplified map showing bottomlands and terraces on the northwest side of the Shenandoah Valley, Va., to illustrate the formation of pediments by lateral planation and piracy.

and the North River have their headwaters in an area of resistant rock of Silurian, Devonian and Mississippian age consisting mostly of sandstone, quartzite, and shale. The relief in this area averages 1,500 to 2,000 feet and ridges rise to altitudes as high as 4,400 feet. The soft rock area into which the three streams flow is underlain by Cambrian and Ordovician carbonate rocks and shale and has an average relief of only a few hundred feet. Most of it is less than 1800 feet in altitude. On entering the carbonate rocks each of the three streams is bordered by a broad floodplain and terraces composed of cobbles and sand derived from the mountains upstream. The low hilly plain at the foot of the mountains is an extensive complex of dissected terrace remnants that resembles, and in fact is similar to the dissected pediments common at the foot of mountain ranges in the western United States.

Mossy Creek, a tributary of the North River, does not head in the sandstone mountains, but in the limestone area. It has a flatter gradient than the North River even though a smaller stream and at locality A (fig. 4) where the two streams are separated by a low divide, Mossy Creek is about 60 feet below

the elevation of the North River floodplain. In 1949 during a severe flood, water actually spilled over the divide into Mossy Creek. This is therefore an example of a stream piracy in progress.

In the resistant rock area the relief is high because the bedrock is removed largely through mechanical processes. Slopes are steep, divides sharp and there are many rock slides. Stream slopes throughout are adjusted for any given drainage area and discharge to transport rock fragments of large size. In the soft rock, or carbonate area, on the other hand, chemical weathering is more important and although the surface is being lowered at the same rate as in the resistant rock area the graded slopes are much gentler and the relief lower. Where the North River leaves the resistant rock area it moves a large load of sandstone cobbles on its bed. They are carried out onto the soft rock area where the channels are adjusted for the transportation of much finer debris. As a consequence the stream shifts laterally and deposits cobbles on the banks, forming a floodplain. Being more resistant to weathering than the carbonate rocks on which they have been deposited, the cobbles persist in the landscape for long periods and as the river continues to erode they form terraces and eventually cap divides.

Mossy Creek will continue to erode its channel and since it is not required to move a load of cobbles, momentarily cuts faster than the North River. Eventually piracy will occur and Mossy Creek Valley will be aggraded by cobbles brought in by the North River. The North River floodplain below locality A will then be abandoned and will become a dissected terrace. Captures similar to this one have already occurred at other places. Note for example in figure 4 the low terraces that connect Briery Branch with the North River.

In this explanation it has been assumed that the resistant and soft rock areas are in the same erosional system and that the average rate of erosion in the two parts is the same. The pediment area exists because cobbles are shed from one part to the other, thus introducing and maintaining a belt of resistant cobbles at the margin of the soft rock area. The pediment area is of course itself in equilibrium in the same system. Its size and the amount of relief are determined by the rate at which cobbles are carried into the soft rock area (a function of size of drainage basin) and by the rate at which they are weathered and broken into pebbles that can be moved out in the streams draining the soft rock area. Hunt, Averitt, and Miller (1953, p. 189) applied the same kind of explanation to the pediments in the Henry Mountains. In that area, however, because the climate is semi-arid the processes are not quite the same, for there is a loss of discharge involved as the streams leave the more humid mountain area.

In the Valley and Ridge Province of the Appalachian Highlands there are many extensive surfaces produced by the processes just described. Many valleys are floored by shale and bordered by relatively high sandstone ridges. The gravels shed by the ridges are too coarse to be carried off by the master stream flowing in shale, and so are stored in floodplains, terraces, and dissected terraces, that because of their resistance to erosion form high benches on either side of the master streams. Eventually the cobbles and gravels in these benches

are reduced by weathering and reworking in the laterally shifting streams to sizes that can be carried off down the main valley. Such extensive pediment-like landscapes have long been mistaken for a former gravel-covered broad valley stage or peneplain that is now dissected. Actually such gravelly surfaces testify to the contrast in resistance between the rocks of the mountains and the rocks of the valley and they are part of the equilibrium between the two.

Terraces.—Some stream terraces may have their origin as equilibrium forms. Mapping of surficial deposits in the Shenandoah Valley, Va., indicated that terraces are most common in soft rock areas along streams that originate in hard rock areas. This coincidence suggests that the terraces are preserved essentially because they contain components more resistant than the underlying rock. Terraces composed of chert cobbles are common in areas of cherty limestone. They are not common, however, in areas of homogeneous rocks of any kind that do not provide a possibility for a contrast in resistance between the stream load and the rock through which it moves.

THE RETREATING ESCARPMENT AND PARALLEL RETREAT OF SLOPES

According to cyclic theories elaborated by Penck (1953) and King (1953) the evolution of topographic forms involves parallel retreat of slopes. As the hill or mountain slope retreats away from the main stream a pediment or network of interconnected pediments forms at the foot of the retreating slope and is extended as erosion continues. The Badlands area of South Dakota provides an example (Smith, 1958). Such forms are, however, far from universal. The retreating escarpment appears to be characteristic of gently dipping stratified rocks in which some beds are more resistant to erosion than others. Pediments and foot slopes may or may not be associated with them. The Highland Rim of Tennessee is an example of an escarpment that may be retreating but is without pediments at its foot. The Badlands of South Dakota, on the other hand, have in front of them a broad network of miniature pediments.

Retreating escarpments and associated pediments appear to be especially characteristic of dry climates, a relationship pointed out by Frye (1959). They are certainly less common features, however, in the Appalachian Highlands and in the Piedmont of the eastern United States where not only is the climate humid but horizontally bedded rocks are rare. In these areas the escarpments are fixed in space by geologic contacts. There may, however, be exceptions. The Blue Ridge escarpment of Virginia and the Carolinas appears to divide rock areas that are in some places identical. This scarp may indicate a condition of disequilibrium between two areas (Davis, 1903; White, 1950; Dietrich, 1958).

EVOLUTION OF TOPOGRAPHIC FORMS THROUGH TIME

The theory of dynamic equilibrium explains topographic forms and the differences between them in a manner that may be said to be independent of time. The theory is concerned with the relations between rocks and processes as they exist in space. The forms can change only as the energy applied to the system changes. It is obvious, however, that erosional energy changes through time and hence forms must change. It is of interest, therefore, to speculate on

the effect of a gradual reduction in relief of a well-graded landscape such as we assume occurs through long periods of geologic time as diastrophic forces cease to exert their influence and an isostatic balance is approached.

In a typical ridge and ravine landscape the general character of the topography is probably maintained as the relief is lowered. There is no reason to believe that the efficiency of the forms for the shedding of waste becomes any less. The forms in which the waste is removed may change, however, and the rate of removal may diminish. In an area of high relief the waste may be largely in the form of boulders and cobbles that are removed mechanically. As relief is lowered in the same area, perhaps chemical weathering becomes relatively more important. In a high relief area, the divides are sharp and slopes steep. As relief is lowered in the same area the slopes in interstream areas become more rounded, and the divides more blunt.

Speculating on the evolution of pediment landscapes that occur in soft-rock areas adjacent to hard-rock areas it is evident that if relief becomes lower the difference in the energy potential between the two areas will become less marked. It can be expected therefore that the pediment areas will diminish in size and may eventually disappear.

THE APPLICATION OF GEOMORPHIC CONCEPTS TO GENERAL GEOLOGIC PROBLEMS

The theory of the geographic cycle has been widely used by geologists. Its abandonment must result eventually in changes in many of our concepts. Though it is not my purpose here to elaborate fully such changes, some examples are cited in order to illustrate the extent to which a change in theory may affect geologic problems.

Theories of ore genesis, particularly of deposits classed as supergene, are affected by abandonment of the cyclic concept of landscape evolution. The manganese deposits of the Appalachians associated with lower Cambrian carbonate rocks are a good example. These deposits generally occur in thick residuum preserved beneath quartzite gravels shed from adjacent highland areas. They have been interpreted as of Tertiary age, formed in the Harrisburg cycle of erosion (Hewett, 1916, p. 43-47). By application of the theory of the equilibrium landscape they may be interpreted not as relics of a Tertiary weathered mantle preserved beneath younger gravels, but as deposits that are forming at the present time, or under conditions like the present: They form beneath the gravelly mantle, and are preserved because the gravel covers them, and protects both the ore minerals and the residuum around them from erosion (Hack, 1959).

Some of the greatest changes in concept required relate to the concept of the dissected peneplain. Because we have accepted for many years the idea that a ridge and ravine landscape is formed by the dissection of a peneplain we have also accepted the idea that many highland areas like the Appalachians eroded in steps or cycles and that the orogenies that deformed the rocks of such highland belts were followed by long periods of vertical uplift of a cyclic nature involving repeated changes in the rates of deformation. Having abandoned the peneplain we must reexamine the history of such areas and apply areal

studies of erosional process and form to the interpretation of their past history. In the Appalachian Highlands, for example, the general outlines of the present drainage may be inherited in part from conditions that existed as early as Permian or Triassic time. The present landscape may have formed through one continuous period of dying orogeny or isostatic adjustment. Differences in relief and form in different areas are explainable partly by the reaction of various erosive processes on a complex bedrock, and partly by what is probably a long history of complicated diastrophic movements.

Cyclic theories of landscape origin are close relatives of the theory of periodic diastrophism which holds that orogenies have generally occurred in geologic time in brief episodes of world wide extent. This theory is questioned and critically discussed by Gilluly (1949) who shows that the evidence of the sedimentary rock column supports the idea that diastrophism has not been periodic but was almost continuous through time, though the form and location of diastrophic movements has continually changed. This concept of continuity of diastrophic processes is, of course, discordant with cyclic geomorphic theories, but is in harmony with the equilibrium concept outlined here.

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ON EVOLUTIONARY EURYHALINITY

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ABSTRACT. The Monera and many of the lower groups of Protista exhibit extreme evolutionary euryhalinity and it is impossible to state whether the lowest groups of organisms are characteristically inhabitants of fresh or salt water. In the Metazoa, the capacity to invade freshwater from the ocean is very irregular and probably represents a statistically superdispersed system. Such a system could be explained formally by supposing that the various physiological preadaptations required to invade fresh waters are distributed with characteristic but different probabilities. Once the osmotic problems have been surmounted by an invader, the problem of salt uptake from very dilute solutions remains. It is probable that many animals of marine origin which are found associated with freshwater species only in coastal localities have simply failed to develop really efficient salt uptake mechanisms, even though they are osmotically well adapted to freshwater. There is probably some correlation between small size and evolutionary capacity to change from one environment to the other. The evidence from the Caspian basin suggests that chemical adaptation is only one aspect of invasion into freshwater; it is suggested that the preexisting freshwater fauna now constitutes the major barrier against invasions from the sea.

INTRODUCTION

The following conclusions have emerged from a comprehensive survey of our knowledge of the origins of the freshwater biota, undertaken during the preparation of the second volume of my *Treatise on Limnology*. These conclusions are presented in a relatively undocumented form at the present time, in honor of one of the great investigators of paleolimnology, and in the hope that they may be of interest to paleontologists and others concerned with the historical problems of the biosphere.

THE CONCEPT OF EVOLUTIONARY EURYHALINITY

A taxon will be said to exhibit evolutionary euryhalinity at its particular hierarchical level, if it includes at the next lowest level members living in both the ocean and in freshwaters.

Obviously the most extreme type of euryhalinity, not however conforming to the definition, will be the euryhalinity, in the ordinary sense of the word, of an individual which can be adapted to either environment. In such a case no genetic change is involved. In certain of the lower organisms it appears that one or a few mutations, well within the specific limits of variation are all that is needed to produce the requisite evolutionary change. In many genera (e.g. *Chlamydomonas*, *Asterionella*, *Melosira*, *Brachionus*, *Plagiostoma*, the minute interstitial crustacea, etc.) there are both marine and freshwater species, so that the change in environment occupied has been accompanied by genetic change involving not more than evolution of new species. In some families (e.g. Desmomonadinaceae, various lower flatworms), marine and freshwater genera occur, indicating a weaker type of evolutionary euryhalinity involving evolution of new genera. As one considers progressively higher taxa, the probability of evolutionary euryhalinity increases, but its interest decreases. There are, however, some phyla, such as the echinoderms and brachiopods, that do not exhibit the phenomenon.

DISTRIBUTION OF EVOLUTIONARY EURYHALINITY IN THE LOWER ORGANISMS

In the Monera (Eubacteriae, Spirochaetae, Myxophyceae, Myxobacteriae), evolutionary euryhalinity is common at the generic, specific, or infra-specific

level. In the marine bacteria (ZoBell, 1946; Roach and Silvey, 1959; Pratt and Waddell, 1959) there is evidence that marine forms on continuous cultivation may become more euryhaline, and it seems likely that in many groups the amount of genetic change needed to convert a marine bacterium into one living in freshwater is very small. Most blue-green algae occur in inland waters, but there are many species that exhibit great tolerance, particularly among the simpler forms, both coccoid and filamentous.

Among the Protista the same situation is found among the lowest forms as prevails among the Monera, while in some of the more specialized groups there is a marked restriction to either marine or freshwater habitats. This is very well shown by the brown algae, in which freshwater as well as marine forms occur only in the small and primitive Ectocarpales, all the higher forms being marine. A comparable, if slightly less diagrammatic, pattern is provided by the red algae. In the green algae extreme evolutionary euryhalinity is exhibited in the lower unicellular Volvocales, while the higher orders tend to settle down into marine (e.g. Siphonales), or freshwater (e.g. Conjugatae and Charales) groups. Similarly the more primitive genera or families of rhizopod and actinopod protists are much less definitely marine or freshwater groups than are the Foraminifera, Acantharia or Radiolaria. Some quite specialized groups, however, seem to retain a marked degree of evolutionary euryhalinity, notably the diatoms, dinoflagellates, heliozoa and ciliates. The resulting picture is best generalized as a progressive loss of evolutionary euryhalinity among the specialized forms, but to a varying degree. Such a picture lends no support to the familiar dictum that life began in the sea.

EVOLUTIONARY EURYHALINITY IN HIGHER ORGANISMS

The higher freshwater plants are clearly of terrestrial origin and do not concern us. Although the same is true of some of the most specialized freshwater animals, such as the aquatic Hemiptera, the freshwater metazoan fauna as a whole is ultimately derived from the ocean. The only possible exception is in the lower flatworms, which perhaps retain the same sort of evolutionary euryhalinity as is exhibited by the ciliates. If the flatworms are derived from ciliates by internal differentiation of cells around numerous macronuclei (see Hanson, 1958 for a recent review), or if the ciliates are really degenerate flatworms as Cain (1959) has suggested, this comparison may be of some interest; but even accepting a phylogenetic relationship between the groups, it must be remembered that the existing acoels and related very primitive forms such as *Xenoturbella* and *Nemertoderma* appear to be strictly marine.

In all the other phyla of animals possessing freshwater as well as marine representatives, the distribution of the former in the taxonomic system is highly irregular, suggesting a great degree of superdispersion of the physiological characters that pre-adapt marine organisms to entrance into freshwaters.

Thus in the sponges the only freshwater families, namely the *Spongillidae* of wide distribution and the *Lubomerskiidae* in Lake Baikal, with probably allied genera *Metschnikovia* in the Caspian and *Ochridaspongia* in Lake Ohrid, are both allied to the marine *Haliclonidae*. In the Coelenterates, with an enormous marine littoral fauna, only four invasions appear necessary to produce

the freshwater fauna. The *Hydridae* are presumably of littoral marine origin; *Cordylophora*, probably not fully adapted, belongs to the *Clavidae*; *Podypodium* is a narcomedusan and the freshwater medusae belong to the *Olindiidae*. The last named family is close to the *Moerisiidae*, which are mainly brackish. A comparable pattern is found in the polychaets (Hartman, 1938, 1951) in which nearly all the fully adapted freshwater species belong either to the *Fabriciinae* (*Sabellidae*) or to the *Nereidae*, particularly to the group of genera near *Namanereis* (=*Lycastis*). Study of other cases, in the Mollusca or Crustacea, where a large freshwater fauna has developed, indicates that this fauna has arisen in a similar irregular way, and that in the origin of any given line, it is hard to say if a single invasion, or a group of invasions by closely allied forms, occurred.

SYSTEMATIC ASPECTS OF THE PHYSIOLOGICAL PROCESSES INVOLVED

It is evident that for a metazoan to invade freshwaters from the sea, several specific conditions must be met. If these depend on various physiological preadaptations randomly distributed throughout the animal kingdom, each with its own frequency, the very superdispersed appearance of the incidence of invasion might have a formal explanation. Unfortunately, until the apparent superdispersion can be treated quantitatively, such a possible explanation is of little value.

The first physiological requirement is presumably that the tissues of a potential invader must be able to adapt to operating at a lower electrolyte concentration than that usual in many organisms. There are obvious limits to the permissible dilution but these limits apparently vary from group to group.

Secondly, some degree of osmotic regulation in the dilute medium is essential to meet the limits set by the requirements of the tissues. This involves retention of salts and exclusion or excretion of water. Exclusion of water across respiratory surfaces can never be complete.

Thirdly, to permit growth, a salt-absorbing mechanism is necessary even if all the salt in the organism can be retained, which is probably never quite possible. In most cases since the primary plant food in freshwater is very poor in NaCl, a special sodium and chloride absorbing mechanism operating from a very dilute source is needed.

Potts (1954) has shown clearly that the first and second conditions can be met in two extreme ways. The tissues can become adapted to great dilution internally and relative to circulating body fluids. The osmotic intake of water is thus minimized. This is probably the only method available to an organism with a very large epithelial surface, such as the lamellibranch ctenidium, exposed to the external medium. Alternatively, the incoming stream of water may be regulated by reducing, not the intensity factor, but rather the capacity factor, most of the surface of the organism being made as impermeable as possible so inflow occurs across limited areas. This is probably the usual situation in the Crustacea, in which the blood and tissue electrolytes are retained at a much higher concentration than in freshwater clams.

A rough count of the number of invasions needed to produce the modern fully adapted freshwater fauna indicates about forty for the Crustacea, and

between twenty and thirty for the Mollusca. Since the latter group is considerably more numerous in species in the sea than is the former, it is doubtless safe to conclude that the probability of any given Crustacean line entering freshwater from the sea is over twice as great as for any given Molluscan line. This is in accord with intuition based on Potts' physiological conclusions.

THE SIGNIFICANCE OF SMALL SIZE

Outside the Monera and the smaller Protista, the groups that seem to exhibit most evolutionary euryhalinity at the generic or lower level are either the fishes in which very special osmoregulatory mechanisms exist, or the various smaller free-living turbellaria, the rotifers and those minute Crustacea of the interstitial fauna (*Microparasellinae* and *Microcerberinae* in the Isopoda; *Bogidiellidae* and *Ingolfiellidae* in the Amphipoda) whose discovery has been one of the more interesting advances in systematic zoology in the past few decades. One is naturally impelled to ask whether among the invertebrates which may be assumed poikilosmotic in the sea, small size *per se* may not be one condition for easy evolution of freshwater from saltwater species. A positive answer would be physiologically surprising, for it would be reasonable to suspect exactly the reverse situation; such rather indefinite data as exist, however do seem to suggest that this factor may operate even if in a presently unknown way.

THE PREVALENCE OF IMPERFECT ADAPTATION

It is very frequent, particularly in Crustacea, to find species that appear to be the associates of normal freshwater animals, but with a strictly coastal distribution. Some of the best examples are found in the Mysidacea. This group contains about four hundred species, of which about twenty-five are known in water containing less than 0.5 percent salinity (Stammer, 1936; Banner, 1953), and so may be supposed to have normal freshwater associates. In spite of these rather impressive figures, the greater number of the supposedly freshwater Mysidacea have limited coastal distributions; the only widespread truly freshwater species are *Mysis relicta* in the regions connected with proglacial late Pleistocene lakes, two or three *Paramysis* in the Volga basin and perhaps the euryhaline *Gangemysis assimilis* in the Ganges. The various species of *Neomysis*, *Mesopodopsis* and *Diamysis* recorded from supposedly freshwater, and the peculiar *Taphromysis* from Louisiana and *Nanomysis* from Siam, seem to be coastal, while a few further forms have been isolated in caves. The group clearly makes a good beginning in invading freshwaters, but contributes very little in the end. The same is true of various other groups of Peracarida. While part of the difficulty is doubtless due to competition, it is reasonable to suppose, in line with the work of Krogh (1929) and Wikgren (1953) on other freshwater animals, that the final physiological adaptation, to water containing less than say 10 mg Cl per liter or about 0.002 percent sodium chloride salinity, can only occur with the development of a really efficient salt uptake mechanism. Coastal freshwaters, enriched with NaCl from sea spray, would support less efficient salt absorbers than would inland freshwaters, but the investigator who relied on his tongue as an analytical instrument would class both sets of waters as fresh.

RELATIONSHIP OF SALT ABSORPTION TO DEVELOPMENT

Needham (1930) pointed out that the rather general suppression of free swimming larvae in freshwater animals might be due not so much to the dangers of such stages being swept down rivers as to the necessity of providing large mineral reserves in freshwater eggs. Passage from salt to freshwater would therefore be facilitated by the development of larger and thus commonly fewer eggs, hatching at an advanced stage of development. Data bearing on this conjecture are severely limited, but it is interesting to find that Behning (1925) observed, in several species of Caspian Crustacea that have invaded the Volga, an increase in egg-number in the fluviatile populations. It is probable that this case represents adaptation to fluviatile conditions without any significant variation in chemical parameters. In so far as the data go, therefore, they support Needham's view, which would imply that on entering a river from an equally dilute sea, no increase in size and consequent decrease in egg number should occur. It is indeed possible that other things being equal, a fluviatile environment in which many young may be carried down stream, might select for numerous eggs, even if this implied a smaller egg size. Unfortunately no measurement of the eggs was given.

COMPETITION AS THE MAIN LIMITATION TO THE EVOLUTION OF NEW FRESHWATER ANIMALS

The Caspian Basin probably has provided the most favorable existing conditions for the evolution of freshwater animals. It is not impossible that during the later parts of glacial periods when there was a connection with the Pontic Basin by the Manych Channel, particularly at the time of the cutting of the Dardanelles, the whole system was filled with freshwater. In spite of the favorable nature of the chemical conditions and the very large number of amphipods, Mysidacea and Cumacea that still live in the northern Caspian under practically freshwater conditions, the amount of invasion into the western Eurasiatic continent is small. Behning (1924, 1925) lists forty-four Peracarida from the Volga delta where the water is essentially fresh. Of these only the fluviatile subspecies *Corophium curvispinum devium* has been able to invade western Europe from the basins that drain into the Ponto-Caspian seas, while two mysids and two gammarids have achieved a very wide distribution in the Volga and other rivers running into the Ponto-Caspian system. Thus of the species that must be almost perfectly chemically adapted, only about 11 percent have been able to achieve significant ranges as freshwater animals. A comparable picture is presented by the diversified molluscan fauna, which has only added *Driessena polymorpha* to the freshwater fauna of western Eurasia as a whole. It is reasonable to suppose that at least one factor of major importance is that the invasion is into a region that already has a well developed freshwater fauna.

That this is a reasonable interpretation is shown by the extreme diversity of the hypogean freshwater fauna which includes numerous animals of marine affinities, clearly stranded in gradually freshened coastal systems of caves, or invaders of interstitial waters. Many of these animals have no close freshwater relatives. The fact that they can adapt to freshwater indicates that their marine

ancestors were potential invaders, but could only enter under circumstances in which little or no competition from existing freshwater animals occurred.

ACKNOWLEDGMENTS

I am indebted to Dr. Peter Gray for calling my attention to the extreme euryhalinity of some marine rotifers, and to Dr. E. S. Deevey for reading the present manuscript.

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PROBLEMS OF STRATIGRAPHY AND CORRELATION OF PRECAMBRIAN ROCKS WITH PARTICULAR REFERENCE TO THE LAKE SUPERIOR REGION*

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ABSTRACT. Two great advances in geologic knowledge—the gradually accruing evidence of the immense duration of Precambrian time, which is now believed to represent about 8 times that of the Paleozoic, Mesozoic, and Cenozoic combined, and development of the facies concept, with concomitant destruction of "layer cake" stratigraphy—have forced thorough reappraisal of stratigraphic correlations of Precambrian rocks. This reappraisal, which is being made by many students of the Precambrian, has thrown doubt on heretofore accepted correlations and has led to greater caution in regional syntheses. Most of the familiar time and time-stratigraphic terms have been virtually abandoned except in type areas, but despite this apparent retrogression, Precambrian stratigraphy probably is on a sounder basis than ever before.

Within individual areas, even areas of highly deformed and metamorphosed rocks, problems of stratigraphy and of correlation are being solved by detailed mapping and application of the physical criteria of correlation. In the Lake Superior region, for example, about 100 formations, plus named stratigraphic equivalents, can be placed in sequential position: these formations represent a probable time span of more than two billion years. District-to-district correlations are made on the basis of structural position, highly distinctive sequences and "carry-over" units, and bracketing by isotope age determinations on older and younger crystalline rocks. The resultant stratigraphic classifications and correlations cannot attain the refinement possible in fossiliferous strata, but they provide useful approximations for reconstruction of regional geologic history.

INTRODUCTION

> The history of attempts at correlation (of Precambrian rocks) has been one of early simple generalization, followed by the recognition of many qualifications and increasing complexity to a point where the initial simple assumptions become almost unrecognizable or are abandoned entirely. Looking back on most of the sweeping generalizations which were made in the past, it is easy to recognize the influence of the Wernerian hypothesis of sedimentation, the assumption of uniform marine conditions, the assumption of uniform orogenic and epeirogenic disturbances over wide areas, belief in the wide extent and uniformity of unconformities and metamorphic conditions, and, finally, a lingering faith in the existence of an igneous basement complex, possibly a remnant of the primordial shell, affording a world-wide base for beginning pre-Cambrian sedimentation. (Leith, 1934, p. 167).

This statement, made by Leith in his presidential address to the Geological Society of America a quarter century ago, is as pertinent today as it was then. The retreat from the broad generalizations and correlations of the past still continues. By now, most of our familiar time and time-stratigraphic terms—such as Archean, Proterozoic, Algonkian, Laurentian, Keewatin, Algoman, and Huronian—have been abandoned or are under attack except in their type areas; in their place we see a chaotic proliferation of local names that do not lend themselves to regional synthesis. In view of the unquestioned need, both economic and scientific, for syntheses broader than those of small districts, it is worthwhile to look into the basic causes of the apparent degeneration of stratigraphic concepts for the Precambrian, then to consider what can be done and, for one small region, what is being done.

DURATION OF PRECAMBRIAN TIME

The first great factor affecting the status of Precambrian stratigraphy is the immense change that has taken place in the concept of duration of Pre-

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cambrian time. Walcott (1893, p. 675), using rate and volume of sedimentation as a guide, held that the span of Earth's history was about 55 million years, and Sollas (1909, p. cxii), following Joly, recomputed the salt content of the oceans to arrive at an estimate of "between 80 and 150 millions of years". In 1917, the classic paper by Barrell was published in which full account was taken of radiometric age determinations (Barrell, 1917), but even as late as 1924, Clarke, author of the authoritative Data of Geochemistry, doubted the validity of the radiometric method and stated (Clarke, 1924, p. 323):

> From chemical denudation, from paleontological evidence, and from astronomical data the age has been fixed with a noteworthy degree of concordance at something between 50 and 150 millions of years. The high values found by radioactive measurements are therefore to be suspected until the discrepancies shall have been explained.

Barrell's assignments of time for the Paleozoic, Mesozoic and Cenozoic are not greatly different from those now accepted, but Precambrian time was assumed to be "perhaps a billion years" (Barrell, 1917, p. 895). In 1947, Holmes published his "A" and "B" scales, the latter of which has received wide acceptance. Holmes' estimate (p. 147) at that time for the age of the earth was "about 3,000 m.y." Recent work has suggested some appreciable modification of the "B" scale for Paleozoic and younger time units, and most present estimates for the age of the earth are 4,500 m.y. or greater (Patterson, Tilton, and Inghram, 1955). These changes in age concepts are shown graphically in figure 1. The oldest rocks dated by radiometric methods are older than 3,000 m.y. (see

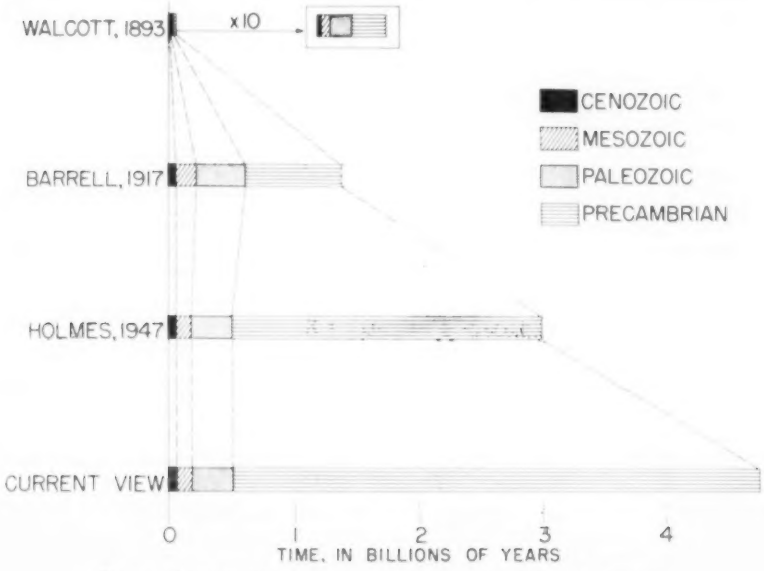

Fig. 1. Graphic illustration of changes in concepts of geologic time.

Aldrich and Wetherill, 1958, for a recent comprehensive review of radiometric dating).

Of more critical importance to the problems of Precambrian correlations than absolute age assignments, however, are the early underestimates of the *relative* length of Precambrian time as compared to that of younger eras. When much of the major nomenclature of the Precambrian was set up 50 or 60 years ago, the duration of Precambrian time was assumed to be roughly equal to or less than that of the Paleozoic, Mesozoic, and Cenozoic combined. Correlations, under this concept, had a far greater *a priori* chance of being correct than in the light of present knowledge—if separated areas, for example, each contained three well-defined rock sequences separated by profound unconformities, it would not seem unreasonable to conclude that they were correlative. But now that it is known that at least 2,500 m.y. of Precambrian history is represented by rocks—a period at least five times as long as all succeeding time—it is quite evident that the probability of *any* correlation being correct is greatly reduced. To put this in another way: if a Precambrian system equivalent in time duration to the Cretaceous is given a length of 3 inches, 50 years ago it had to be placed in a scale 20 inches long; now this 3 inches has to be given a position in a scale with a minimum length of 9 feet and a possible length of 15 feet.

The significance of the great expansion of the time scale for the Precambrian is still not fully appreciated—probably not even by those of us working actively in the field—and the impact is not restricted to stratigraphy alone. Gilluly (1949, 1950) has effectively pointed out that if correlated sequences in separated areas are not truly equivalent in age, if doubt is placed on equivalence of their bounding unconformities (which represent orogenic intervals), and if proper account is taken of greater opportunity for observation for younger rocks, then some common generalizations, such as those dealing with "increasing tempo of orogenies with time" and progressively greater rate of sedimentation with time, rest on very shaky foundations. In the Precambrian, as for younger rocks, each rejected correlation may add significantly to the cumulative total of stratigraphic thicknesses (and therefore increase the apparent rate—though not necessarily the volume—of sedimentation per unit of time), and it may add one or more orogenies to the record. The ultimate dependence of concepts of this sort upon correlation is perhaps more evident for younger rocks, but it is no less true for the Precambrian. And it is within the Precambrian that correlations are most uncertain and for which unrealistic correlations are clung to most tenaciously. Stille (1955, p. 188-189), for example, writes:

> Between Huronian and Algonkian[1] occurred one of the mightiest folding deformations of the earth, known as the Algoman revolution in American usage or the Karelian revolution elsewhere. One can say that this event produced an essentially complete consolidation of the earth's crust in the sense that there were no regions of appreciable size capable of sustaining Alpine-type folding. *This is borne out by the fact that not a single instance of unquestionable orthogeosynclinal continuity between the Huronian and the Algonkian has been discovered.* (My italics.)

[1] Stille's use of Huronian, Algonkian, and Algoman is, of course, completely at variance with "classic" terminology in the type areas from which these names are derived. As that terminology is summarized by Leith (1935, table opp. p. 10), Huronian is a subdivision of Algonkian and rocks assigned to it rest unconformably on Algoman granite.

Similarly, Bucher (1933, p. 418) states:

> So far as the writer knows, the Archezoic rocks were strongly deformed everywhere on the face of the earth before the Proterozoic era began.

But the argument is completely circular: it is precisely *because* of great structural discordance that rocks in individual regions have been designated Archeozoic and Proterozoic (or "Huronian and Algonkian"). Until the advent of radiometric dating, there was no independent evidence to show, for example, that the Proterozoic of one area is equivalent to that of another, or that their bounding unconformities are the same. Gilluly's trenchant analysis (Gilluly, 1949) shows that even for thoroughly studied fossiliferous strata, the minimum unit of time that can be distinguished is rarely less than several million years, and that events that take place intermittently or continuously during this period are commonly grouped to give an erroneous impression of widespread simultaneity. For the Precambrian, the minimum unit of time that can be discriminated by radiometric methods surely is at least tens of millions of years; as a result, for rocks in excess of a billion years old it is quite unlikely that successive deformations comparable to the Nevadan and Laramide orogenies could be separately defined. It is small wonder that the classic Precambrian deformations appear to have been events of vast significance; each named orogeny, such as Algoman and Laurentian, probably is a composite of many diastrophisms that took place in different places at different times.

The great increase in time assigned to the Precambrian has impact on other facies of the anlysis of earth history. The limitation of the cherty iron-formations to the Precambrian, for example, often has been cited as potent evidence in favor of the view that the Precambrian atmosphere and environment must have been substantially different from those of younger eras. Differences may have in fact existed, but the lack of iron-formation in post-Precambrian rocks is now far from a compelling argument. Iron-formations of many different ages occur in the Lake Superior region. Radiometric dating of rocks in Minnesota (Goldich, Baadsgaard, and Nier, 1957b) indicates that the Biwabik iron-formation of the Mesabi district is 700 m.y. or more younger than the Soudan iron-formation of the Vermilion district. Unless it can be shown that iron formations of intermediate age exist, this could mean that there were intervals within the Precambrian as long as or longer than all of post-Precambrian time during which the specific combination of environmental factors necessary for deposition of iron-formation did not exist.

NEWER CONCEPTS OF SEDIMENTATION AND OROGENESIS

The second great factor in the apparent degeneration of stratigraphic generalizations for Precambrian rocks is the advance in knowledge of spatial and temporal aspects of sedimentation and of orogenesis.

Some part of older views on deposition of sediments is expressed in the term "layer-cake stratigraphy"—the assumption that similar or identical conditions existed over great areas and that the interruptions or changes in conditions of sedimentation were similarly widespread. As a result correlations were acepted because, and often only because, of similarities in lithology; correlations were rejected because of dissimilarities. In the absence of any means of establishing age equivalence or the lack of it, both acceptances and rejections

must be reviewed critically. It is more difficult to establish the existence and nature of lateral changes in unfossiliferous rocks than in fossiliferous rocks, but that these changes must exist is not open to question. The fine study by White of the iron-formation of the Mesabi district (White, 1954) demonstrates that facies analysis of Precambrian stratigraphic units is possible if sufficient data can be acquired.

The development of major terminology for Precambrian strata was also greatly influenced by a concept, largely abandoned by 1900 but still exerting subtle influence, that Precambrian rocks are the products of truly world-wide processes that took place during the transition from a primordial crust to what would be considered normal geologic conditions (for review and criticism, see Irving, 1888). These processes were conceived to be essentially chemical, and they demanded a universal and certain succession; gaps in the record might exist, but the order is invariable. Names such as Laurentian and Huronian were applied to subdivisions and doubtless this accounts in part for the use of these terms in many parts of the world. A curious sidelight of this classification is afforded by the evolution of the word "taconite", now a well-known term for low-grade iron ores of the Lake Superior region. The term stems from "Taconic", one of the major subdivisions of this older classification of the Precambrian, which in turn is derived from the Taconic Mountains of New York, the strata of which are now known to be Paleozoic in age; "taconite" is the only relic of this nomenclature remaining in the Lake Superior region.

Many other correlations were made and names assigned on the basis of similarity in degree of deformation and metamorphism—in fact, the early definitions of "Archean" were in terms of such criteria—and on the assumption that orogenic revolutions were of great extent, perhaps even world wide. The temporal aspects of orogeny are discussed earlier in this essay; the spatial aspects also have significance to the problems of correlation. The concept of orogeny affecting broad areas has long since given way to that of deformations confined to narrow belts, with rapid cross-strike transition from orogenic to non-orogenic environments. Age equivalence between rocks strikingly different in structural and metamorphic characters must not only be accepted, it must be expected. The attempts to define Precambrian orogenic belts for the Canadian Shield (Wilson, 1949) is recognition of the predominantly linear aspect of zones of orogenic deformation. If such belts could be located accurately by geologic mapping, perhaps some headway could be made toward proper correlation of dissimilar rocks.

METHODS OF ATTACK ON CORRELATION PROBLEMS OF THE PRECAMBRIAN

Abundant reasons have been given in the preceding pages for viewing with great skepticism most of the classic generalizations regarding Precambrian stratigraphy and history. How then do we cope with the problems of analyzing and synthesizing the more than three-fourths of all geologic history represented by the non-fossiliferous Precambrian rocks? And cope we must, because no matter what the difficulties, the questions are far too important, for both scientific and economic reasons, to be ignored. Even though we must acknowledge that in some places intense deformation and metamorphism have virtually de-

stroyed original characteristics of the rocks, in many areas and probably in parts of all areas, stratigraphic geology is possible. It is my purpose now to show that in the attack on such areas we are not wholly without weapons, nor are we wholly without success in using them.

The weapons are the physical criteria of correlation and radiometric age determinations.

The most basic geologic law is that of superposition. Though the principle itself can be called obvious, its application to highly deformed rocks is rarely easy. In many Precambrian terranes, the dip of a bed is without significance as to which rocks are younger or older. Determination of the original "top direction" of a bed may not be possible from internal evidence, as it depends first upon whether diagnostic features ever existed and second upon whether they have been preserved. Even where there is preservation, recognition may be difficult, but in rocks of sedimentary origin features such as crossbedding, ripple mark, cut-and-fill, graded bedding, and algal structures commonly are retained to some degree in all but the most intensely sheared and altered terranes. The recognition does of course require knowledge both of the nature of the original feature and the effects of later processes on that feature, as in the familiar example of grain-size reversal in an original graded bed because of metamorphism. In rocks of volcanic origin, one uses relict pillow structures in greenstone and vesicular and breccia tops in subaerial flows. In igneous rocks it may be a pattern of differentiation—granophyre near the top of a gabbroic sill and ultramafic rock near the base, or their metamorphic equivalents. The determination of top direction in areas of low to moderate metamorphism may be based on secondary structural features, particularly drag folds and cleavage, by means of which a particular bed can be located with respect to synclinal or anticlinal fold axes, and by the mapping out of major folds. These procedures are time consuming, and rarely do they yield results that are beyond question, but it is by such means that the stratigraphic order within sequences, and the relationships between sequences, are established for individual districts.

In correlating from one area to another, we must rely on a combination of clues, none of which individually is above suspicion. In districts such as Lake Superior, formations cannot be "walked out" between areas because of lack of exposure or because of structural discontinuities. More often, the gross position of a sequence must be fixed by broader relations to igneous activity, deformation, and metamorphism. This must, of course, be done with full awareness of the dangers involved, but the errors are perhaps more likely to be those of failure to correlate strata that are actually time equivalent than those of incorrect correlation.

Within the shaky framework set by major unconformities, use is then made of distinctive aspects of the succession. This might be a single unit, such as a thick dolomite that might reasonably be inferred to have had wide original extent, but more reliable are distinctive combinations of clastic sediments, chemical sediments, and volcanic rocks. The point can be illustrated by reference to rocks of the Lake Superior region (fig. 2). The Animikie group of Minnesota, for example, consists of the basal Pokegama quartzite, the Biwabik

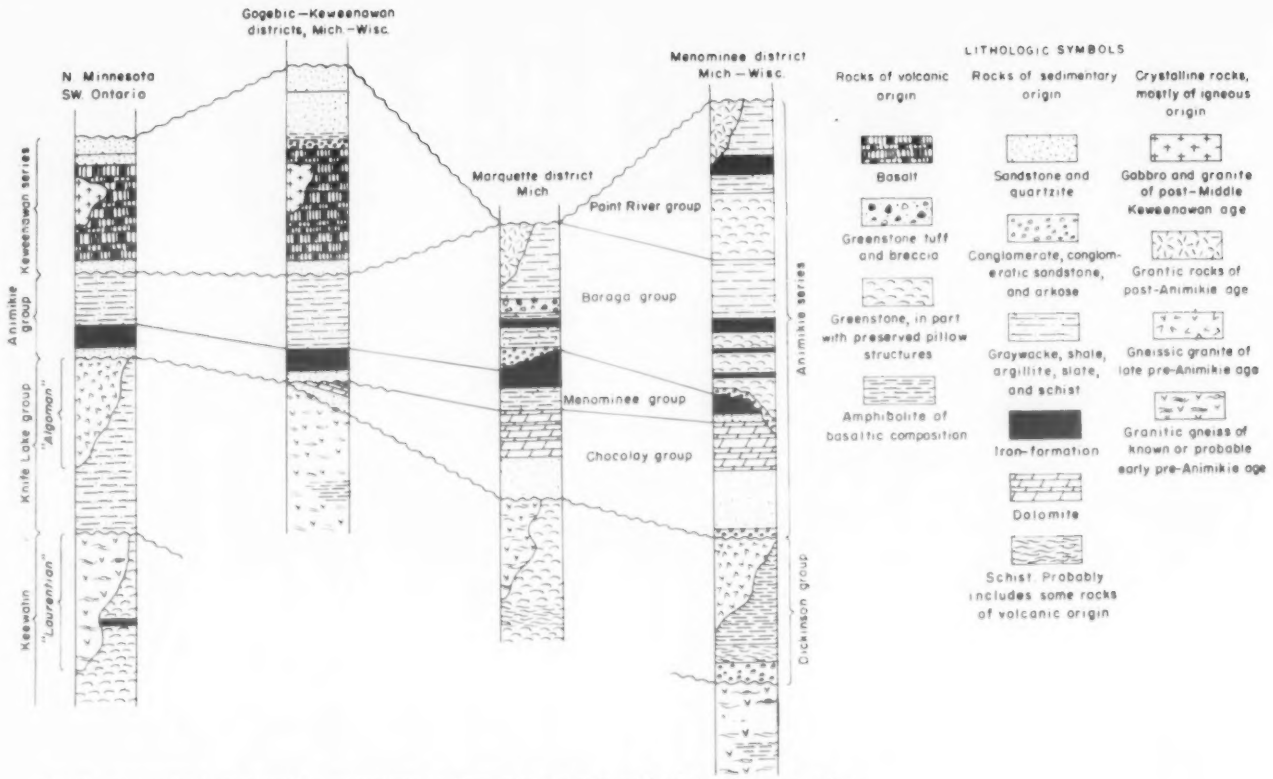

Fig. 2. Generalized stratigraphic successions in Precambrian rocks of the Lake Superior region.

iron-formation, and the overlying Virginia slate[2] (Grout, Gruner, Schwartz, and Thiel, 1951). On the south shore of Lake Superior, in the Gogebic district of Wisconsin and Michigan, a sequence occupying a similar structural position consists of the basal Palms quartzite, the Ironwood iron-formation, and the Tyler slate, which like the Virginia slate is a thick formation of graywacke and argillite. In the Gogebic district, however, remnants of quartzite and dolomite locally separate this sequence from the underlying crystalline rocks. Farther east, in the Marquette and Menominee district of Michigan, the quartzite and dolomite are thick formations that conformably or disconformably underlie the iron-bearing sequence, and the overlying rocks attain great thickness and complexity. The entire succession, despite the presence of minor unconformities, has unity in character in that the rocks provide a record of progressive change in depositional environment from that of a stable shelf, characterized by deposition of quartzite and dolomite, to that of a eugeosyncline characterized by deposition of graywacke and submarine volcanic rocks. It has therefore been designated as the Animikie series (James, 1958). Aside from correlated equivalents and subdivisions of individual formations, it contains about 20 established stratigraphic units that are placed in four groups; the aggregate thickness is at least 50,000 feet.[3]

Below the Animikie group in Minnesota and Ontario are two more groups, the Knife Lake with between 10 and 20 formational units, and the Ely greenstone with the interbedded Soudan iron-formation. Similarly, in Michigan, the Animikie series is underlain by relics of two older groups. But correlation between these older groups is not possible, at least with the means at hand. Overlying the Animikie is the Keweenawan series, which has several major subdivisions and a total thickness of 40,000 to 50,000 feet. In all, at least 100 noncorrelative Precambrian formations, with an aggregate thickness of more than 150,000 feet, can be placed in sequential position within the Lake Superior region. Correlations within this immense pile of rock are of varying degrees of reliability, but still they are adequate for reconstruction of large chapters of geologic history—a history that spans more than 2,000 million years. I might add that there is evidence of life throughout: graphitic rocks in the Soudan iron-formation, algal forms and structures in the Kona dolomite and the Biwabik iron-formation, coal in strata of the Paint River group, and oil in the Keweenawan.

In order to place these sequences in a time scale—and to provide a basis for correlation between separated regions—it is necessary to turn to radiometric dating. While this is the shining hope for ultimate subdivision of the Precambrian, it must not be thought that it requires merely more measurements, or even that ultimate success is assured. In the first place, only in rare cases will it be possible to obtain suitable material of primary or diagenetic origin by means of which the actual age of a stratigraphic unit can be determined; in general, most of the age determinations will be made on minerals of

[2] These individual formations, and others mentioned, are indicated only by lithologic pattern on figure 2.

[3] For nearly 60 years this sequence was called Huronian and correlated with the rocks of the type area 300 miles to the east. The abandonment of that designation is part of the general retreat from older and now uncertain correlations.

igneous or metamorphic origin. The values obtained for the minerals from a metamorphic rock provide only the age of recrystallization, which may be far removed from the original age of the rock unit. In the second place, metamorphic and igneous rocks alike may be remetamorphosed, with varying degrees of response by individual minerals in the rocks, so that the age values obtained may be discordant or lacking in agreement with geologically determined relationships.

The kinds of complexity that must be dealt with are illustrated by the geochronologic data now available for Dickinson County, in northern Michigan. The Felch trough in this area is a complex infold of strata of the Animikie series. These strata rest with profound unconformity on an older granite, which is later than the metasedimentary and metavolcanic rocks of the Dickinson group. Pegmatite dikes cut the granite and older metamorphic rocks, and granite dikes cut the Animikie strata. The relations are shown diagrammatically in figure 3. Age determinations from localities 1-8 were made by L. T.

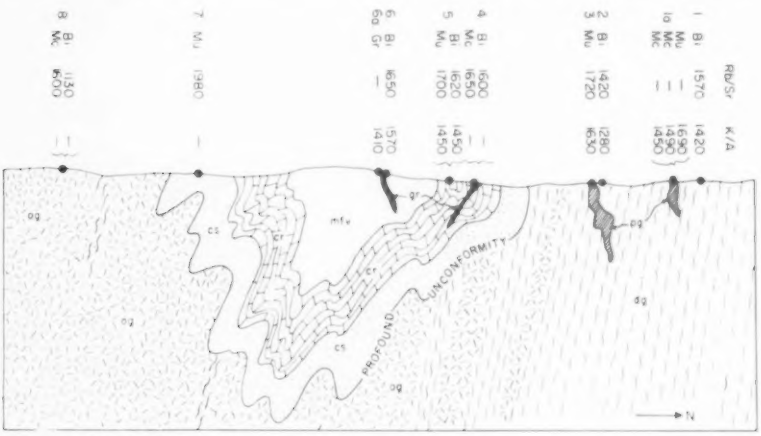

Fig. 3. Schematic cross section of the Felch trough, Dickinson County, Michigan. Rock units as follows: dg, gneiss and schist; og, pre-Animikie granite; cs, Sturgeon quartzite (basal member of Animikie series); cr, Randville dolomite; mfv, Felch formation and Vulcan iron-formation. Pre-Animikie rocks cut by pegmatite (pg); strata of Animikie series cut by granite (gr). Age determinations on biotite (Bi), muscovite (Mu), microcline (Mc) and mica-free granite (Gr), in millions of years; sample localities (solid circles) projected as much as 10 miles into plane of section.

Length of cross section about 5 miles, but units are not drawn to scale.

Aldrich of the Carnegie Institution, using the Rb/Sr and K/A methods, and the data are reproduced here with his permission. Data for localities 1a and 6a are reported by Wasserburg, Hayden, and Jensen (1956, p. 159); these were obtained by the K/A method, with an empirical factor used to correct for assumed argon loss in the microcline and whole-rock samples. The value for the muscovite from locality 1a has been recalculated from the basic data given, using the same constants used by Aldrich. The disagreement between the radiometric age determinations and the geologic history deduced from field

studies is profound, and significant differences exist between ages obtained by different methods or from different minerals in the given rock. This lack of agreements does not necessarily mean that some of the data or geological interpretation must be incorrect, though this possibility must be continually appraised. In actual fact, all may be valid, and each "disagreement" may furnish important evidence leading to a more complete geologic history.

This is not the place to attempt interpretation of the discordant data given in figure 3. A more complete statement of the problem, together with further data and possible solutions, will be made in a forthcoming paper by Aldrich and his associates. But it is clear that any interpretation will call for some of the rocks to have undergone more than one metamorphism.

Two possibilities remain for obtaining an age closer to the real value for remetamorphosed rocks. First, by correlation, a given rock unit may be traced into areas of negligible metamorphism. In Minnesota, for example, where much of the Animikie is very weakly metamorphosed, ages of 2,500-2,700 m.y. have been obtained for minerals in the pre-Animikie rocks (Goldich, Nier, and Baadsgaard, 1957a), and although no correlations can be made with the Michigan rocks at the present time, the possibility does exist. Second, the work of Tilton, Wetherill, Davis, and Hopson (1958) on the Baltimore gneiss has shown that isotopic measurements of uranium-lead in zircon may yield an "original" age whereas K/A and Rb/Sr measurements on micas give an age of a later metamorphism; the "primary" age for the Baltimore gneiss is about 1,100 m:y: and the superimposed metamorphism is about 300 m.y. These methods give hope that most major sequences can eventually be bracketed so that at least crude correlation from region to region or even continent to continent can be eventually made.

In conclusion, I want to make it clear that although the immense duration of time and the lack of diagnostic fossils are formidable obstacles to overcome, the problems of stratigraphy and correlation in the Precambrian can and must be solved. Despite the difficulties, the Precambrian is not a world apart: It contains the same kinds of rocks and reveals the same kinds of geologic processes known from the record of younger eras; the same principles apply and the same rules must be used. And as with rocks of the younger eras, stratigraphy and correlation are the very essence of understanding the geologic record.

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THE ANATOMY AND HABITAT OF LOW-ANGLE THRUST FAULTS*

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ABSTRACT. Low-angle thrust faults, commonly called "overthrusts", are faults with reverse movement, which dip at low angles. Along them, great sheets of rock have moved over the rocks beneath for distances as much as many miles. How they were formed is poorly understood, but much has been learned in recent decades regarding details of their geometry and space relations. Two relatively simple examples, the Pine Mountain fault of the Appalachian region and the Johnnie-Wheeler Pass fault of the Cordilleran region, illustrate this geometry. Initial fractures of these faults, rather than having been regularly dipping shear planes, consisted of a succession of flats and pitches, formed, respectively, on incompetent and competent units of the stratified sequence. This and other characteristics of these examples may also apply to more complex low-angle thrust faults, which have had a more eventful history and have been more obscured by superposed structures. Low-angle thrust faults characterize mainly the miogeosynclinal realm, and in many places seem to have formed near the margins of sedimentary basins or along the boundaries between one sedimentary basin and another.

INTRODUCTION

Through the years, I have made geological surveys in many areas where the rocks have been displaced by low-angle thrust faults. I have also made surveys in some areas where low-angle thrust faults have been proposed, and for which I could find no evidence. The years have increased my understanding of the field expression and the relations in space of these remarkable structures, but the years have brought me little understanding of the manner in which they were formed. It seems mechanically implausible that great sheets of rock could have moved across nearly flat surfaces for appreciable distances (Lawson, 1922, p. 341-342), although recent papers by Rubey and Hubbert (Hubbert and Rubey, 1959; Rubey and Hubbert, 1959) have shed new light on how this might have been accomplished.

I here use the term "low-angle thrust faults" for structures which many geologists call "overthrusts". I believe that terminology of structural features should be primarily descriptive, and to me the term "overthrust" introduces connotations of genesis that are perhaps unwarranted. If some faults are *over*thrusts, probably some others are *under*thrusts, and criteria have been sought by which the two might be distinguished (Lovering, 1932). It would appear, however, that there would be little or no difference in the visible effects, whether the upper block were shoved over the lower, or the lower block were shoved under the upper. The preferred term, "low-angle thrust fault", indicates merely what can be observed—a fault that dips at a low angle, on which there has been reverse movement, so that the upthrown side is on its hanging wall.

More significant than the problem of overthrusting versus underthrusting is the problem of whether thrust sheets moved as a result of forces originating within the earth or moved by surficial forces, sliding on declivities primarily under the influence of gravity.

"Gravity tectonics" may well provide an interpretation for many structural features that have hitherto been inexplicable. Thus, along the Heart Mountain

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fault of northwestern Wyoming, a thin sheet of carbonate rocks was formerly believed to have been thrust over a wide expanse of diverse rocks, after which erosion fragmented the sheet into widely separated remnants, all within a relatively brief interval of Eocene time. Recent work indicates that the observed remnants of the supposed Heart Mountain thrust sheet probably moved as detached slabs down a surface of low declivity, from a source area in the northwest (Pierce, 1957, p. 613-620). Similar movement of detached slabs probably also occurred in the northern Alps, and these slabs are not, as formerly supposed, erosion remnants of vast thrust sheets or nappes that once extended unbroken to their roots in the south (deSitter, 1954, p. 333-334).

It has been proposed, further, that most of the visible structures in one mountain system, the northern Apennines, resulted from wholesale transfer of rock masses by superficial sliding (Merla, 1951, 1957; Migliorini, 1952), and the proponents imply that a similar explanation might be widely applicable to mountain structures elsewhere. A geologist accustomed to more conventional explanations of mountain structure will concede that "gravity tectonics" is a likely accessory to the processes formerly inferred, but will question whether the common low-angle thrust sheets of parts of the Appalachian and Cordilleran regions, whose soles are well beneath the surface, and which are still firmly joined to their roots, could have moved more than incidentally by gravity.

Geological concepts have cycles of popularity, as each generation discovers features or processes which are novel, exciting, and apparently significant for the interpretation of earth history. Somewhat before my day, the concept of ancient glaciations and glacial deposits was in the ascendant, and during each year geologists added new episodes and new areas to the growing roster of ancient glaciations. During my own early years as a geologist, the concept of low-angle thrust faulting was similarly ascendant; these faults were observed at many new places, and they were extended still farther by extrapolation. Today, interest has shifted to such concepts as gravity tectonics, strike-slip faulting, granitization, and the work of turbidity currents.

Ascendency of each concept has been followed by an inevitable reaction, when the more exhuberant claims were questioned or discredited. Some of the alleged glacial deposits are now believed to have been mudflows, landslips, or even quite ordinary conglomerates or fault breccias. Some of the supposed great low-angle thrust faults have disappeared under further scrutiny, or their extent has been greatly diminished. One can anticipate an eventual reaction of the same kind to popular concepts of our own day. These reactions are healthful, as they have produced more rigorous scrutiny of the criteria for the various concepts. Specifically, they have strengthened our confidence in the genuine ancient glacial deposits and the genuine low-angle thrust faults.

REAPPRAISAL OF LOW-ANGLE THRUST FAULTING

Modern examination of the concept of low-angle thrust faulting should include attention to the habitat or environment in which this class of faults might occur. Some of my colleagues have found amazing arrays of low-angle thrust faults in their areas of study, yet others have never encountered such faults at all, even though their areas of study are of comparable structural

complexity. Some geologic realms must have been conducive to the development of low-angle thrust faults, whereas others were not. I have not entirely formulated in my own mind what these realms might be, although some possibilities will be mentioned at the end of this paper.

Reappraisal should also be made of the anatomy of low-angle thrust faults, or their geometry and their space relations. A view widely held earlier was that these faults originated as low-dipping fractures or shears, which broke smoothly upward from the basement through the stratified rocks. Many of these fractures were supposed to have developed after the rocks had been deformed, and cross-sections have been published in which varied rocks and folds of both overriding and overriden blocks are neatly truncated by the plane of a low-angle thrust fault (fig. 1). No convincing explanation was given as to how a clean-cut fracture could have formed in materials of varied structure and competence, nor how a thrust sheet composed of such materials could have moved as a mass with little or no rupture.

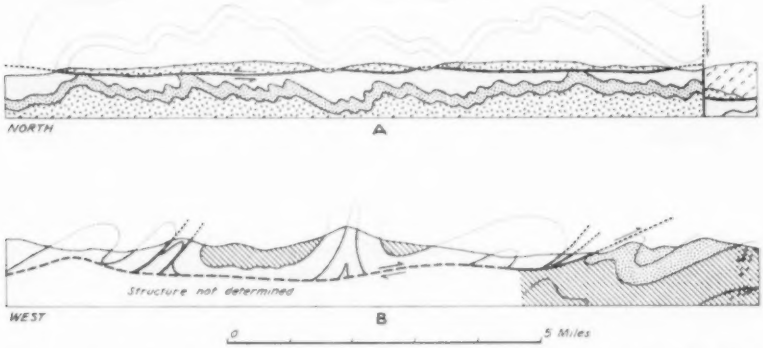

Fig. 1. Low-angle thrust faults which have been interpreted as truncating deformed rocks of the overriding and overriden blocks. A, Reading area, Pennsylvania (after Stose and Jonas, 1935, fig. 8 C, p. 776). B. Bannock fault, southeastern Idaho (after Mansfield, 1927, pl. 12, section T-T'-T'').

PINE MOUNTAIN AND JOHNNIE-WHEELER PASS FAULTS

A different picture emerges where the anatomy of a low-angle thrust fault can be analyzed in detail. Two examples which come to mind are the Pine Mountain fault of Virginia, Kentucky, and Tennessee (Wentworth, 1921, p. 353-358; Butts, 1927; Rich, 1934, Miller and Fuller, 1947; Miller and Brosgé, 1954, p. 187-191), and the Johnnie-Wheeler Pass fault of southern Nevada (Nolan, 1929, p. 465-468; Longwell, 1926, p. 566; 1945, p. 420-423). The anatomy of these faults is readily apparent because it is relatively simple, is well exposed, and, in the case of the Pine Mountain fault, has been penetrated in many places by drilling. The faults have not been obscured, as have so many others of their class, by a multitude of adjacent or superposed structures, or by an eventful subsequent history.

The Pine Mountain fault, where it emerges on the northwest, is a high-angle thrust fault[1], along which a plate of Devonian, Mississippian, and Pennsylvanian rocks is raised against Pennsylvanian rocks (fig. 2A). Southeast of the fault trace are the broad Middlesboro syncline and Powell Valley anticline. The Pine Mountain fault was supposed at first to maintain its steep angle into the depths, and no more than a coincidental relation to the syncline and anticline to the southeast was suspected. In 1926, however, Butts (1927, p. 1) discovered some windows in the core of the anticline, where Cambrian and Ordovician rocks are faulted over Silurian rocks. The fault around the windows could be none other than the Pine Mountain fault, but here it is 17 miles southeast of its main trace, and involves rocks much lower in the stratigraphic sequence.

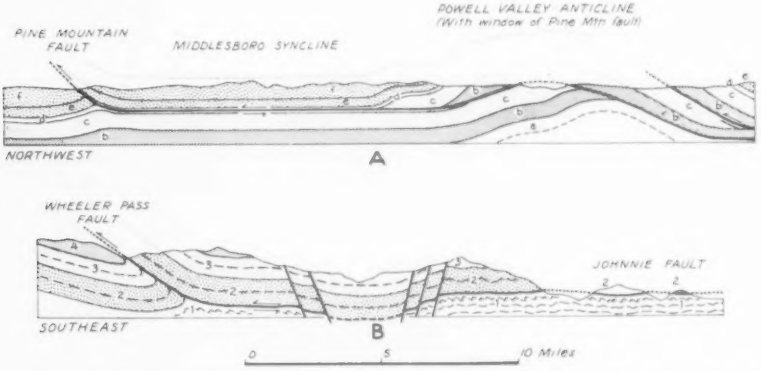

Fig. 2. Low-angle thrust faults whose anatomy is well demonstrated by outcrops and drill data. A. Pine Mountain fault, Virginia, Kentucky, and Tennessee (after Miller and Fuller, 1947, sheet 1, fig. 4). B. Johnnie-Wheeler Pass fault, southern Nevada (compiled from Nolan, 1929, fig. 1, p. 462; Longwell, 1945, fig. 4, p. 422; and Longwell, oral communication). In order to facilitate comparison of the structures, directions are reversed on the two sections. Letters in A indicate a—Lower and Middle Cambrian (Rome and Conasauga formations), b—Upper Cambrian and Lower Ordovician (carbonate rocks), c—Middle and Upper Ordovician and Silurian, d—Devonian (shale), e—Mississippian, f—Pennsylvanian. Numbers in B indicate 1—Johnnie formation and 2—Stirling quartzite (both of Early Cambrian age), 3—younger Cambrian strata, and Ordovician to Mississippian, 4—Pennsylvanian.

Inspired reconstructions by Butts (1927), now confirmed in many places by drilling, indicate that descent of the fault in the sequence is not along a regularly dipping shear plane (fig. 2A). Instead, at the front of the fault on the northwest, it dips steeply and raises older rocks over younger—Devonian shale over Pennsylvanian. Immediately southeast, beneath the Middlesboro syncline, the fault lies in the nearly flat-lying Devonian shale. Recent drilling

[1] Distinctions between "high-angle" and "low-angle" thrust faults are admittedly vague. The two terms could be used empirically for faults which dip, respectively, more than 45° and less than 45°. In this paper it will be more meaningful to use the term "high-angle thrust fault" for one with appreciable dip, which strongly transects the strata in the overridden block, the overriding block, or both. Many faults of this kind dip more steeply than 45°, but some dip at lower angles.

indicates that the shale has been sheared and slightly metamorphosed, and contains a "blow-out zone" in which gas has accumulated (Young, 1957, p. 2568-2672). It is an incompetent layer, which served as a gliding plane along which younger competent Mississippian and Pennsylvanian rocks could override older competent Silurian and Ordovician rocks. Near the windows along the Powell Valley anticline to the southeast, the fault is gently arched and older rocks are again thrust over younger—Cambrian and Ordovician carbonate rocks over Silurian rocks, or rocks a little higher or lower (Miller and Fuller, 1947, sheet 2; Miller and Brosgé, 1954, p. 191-195). On the southeast flank of the Powell Valley anticline the fault descends steeply through the competent Cambrian and Ordovician carbonate rocks, but probably flattens in the next incompetent units beneath, the Conasauga and Rome formations of Middle and Early Cambrian age; here, younger rocks may again be thrust over older rocks.

Drilling has not penetrated these deep-lying roots of the Pine Mountain fault, but an analogous zone is at the surface in southern Nevada (fig. 2B). Within the Spring Mountains the Wheeler Pass fault emerges as a high-angle thrust fault, along which Cambrian rocks are raised over Pennsylvanian (Longwell, 1926, p. 566). Along the west side of the Spring Mountains, about 9 miles away, is the Johnnie fault, along which the Early Cambrian Stirling quartzite overrides the next older Johnnie formation (Nolan, 1929, p. 465-468). The two faults are clearly related, and the whole may be termed the Johnnie-Wheeler Pass fault. Along the Johnnie fault the competent Stirling quartzite is somewhat beveled at its base but is otherwise little deformed. By contrast, the underlying shaly and incompetent Johnnie formation is intensely contorted and phyllitic near the thrust plane, with its deformation and metamorphism decreasing downward. Westward extensions of the Johnnie fault are not exposed, but Nolan (1929, p. 469-471) supposes that it dies out in this direction, with diminishing contrasts in deformation between the rocks above and beneath it.

Relations along the Pine Mountain and the Johnnie-Wheeler Pass faults indicate that their thrust sheets moved over irregular surfaces. The faults ascend over a succession of flats and pitches. The flats occurred where the initial rupture followed an incompetent stratum (the shales of Devonian age and the Rome or Conasauga formations along the Pine Mountain fault, the upper part of the Johnnie formation along the Johnnie-Wheeler Pass fault). The pitches occurred where the initial rupture crossed diagonally through enclosing competent rocks. Some of the diagonal ruptures may have formed by simple shearing across the competent units (Rich, 1934, p. 1589-1590), others may have been localized by early wrinkles or anticlinal folds (Longwell, 1926, p. 568-569; Nolan, 1929, p. 469-471). The number of flats and pitches is conditioned by the arrangement of incompetent and competent strata in the sequence. Along the Pine Mountain fault there are at least two pitches (on the Cambrian and Ordovician carbonate rocks, and on the Mississippian and Pennsylvanian rocks). On the Johnnie-Wheeler Pass fault Longwell (oral communication) believes that the pitch in the rocks above the Johnnie formation extended to the top of the stratified sequence.

During initial stages along the faults considered, younger rocks moved

over older along the flats, older rocks over younger along the pitches. If movement had been sufficiently prolonged, much of the structure would have been converted into a thrust of older rocks over younger, but this was not attained in either of the examples. The Middlesboro syncline and Powell Valley anticline seem to have been produced, not primarily by compressional folding, but by warping of the thrust sheet as it advanced over the flats and pitches of the fault surface (Rich, 1934, p. 1589-1591).

GENERALIZATIONS REGARDING LOW-ANGLE THRUST FAULTS

Although the faults described are rather simple examples, they provide lessons which may be applicable, to some extent, to other and more complex low-angle thrust faults elsewhere. An instructive form of experimentation is to analyze some of the more complex faults by means of these lessons. Hume

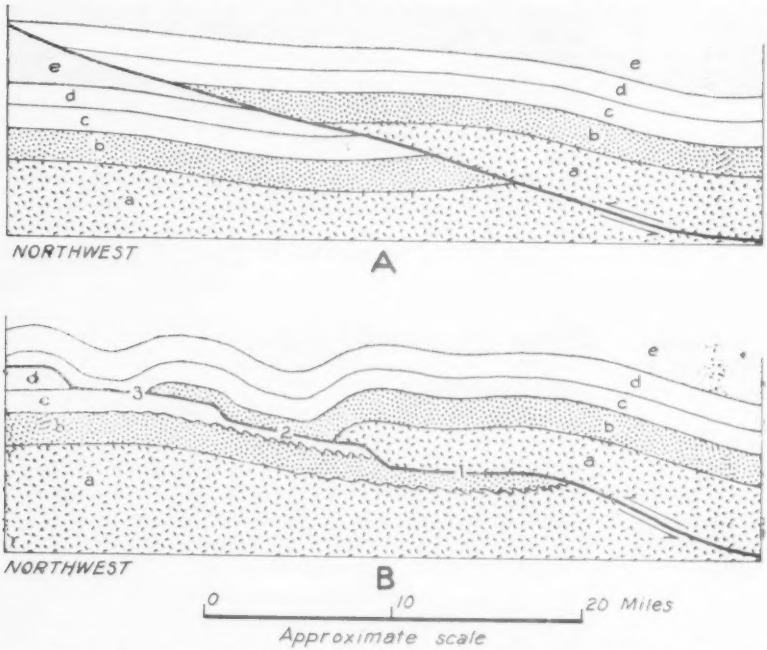

Fig. 3. Interpretations of the form of the low-angle thrust faults of northeastern Tennessee, after thrusting but before later deformation. A. Thrust fault interpreted as a regularly dipping shear plane (Buffalo Mountain fault; after Keith, 1907, fig. 3, p. 9). B. Thrust fault interpreted as a succession of flats and pitches on incompetent and competent strata (Holston Mountain-Iron Mountain fault; compiled from King and Ferguson, 1959). Stratigraphic units are indicated by letters, a—Precambrian basement rocks, b—Lower Cambrian clastic rocks, c—Lower and Middle Cambrian carbonate rocks and shales, d—Upper Cambrian and Lower Ordovician carbonate rocks, e—Middle Ordovician and younger clastic rocks. In B, flats along the thrust fault which developed along incompetent layers are indicated by numbers, 1—in middle of Unicoi formation, 2—in Rome formation, 3—near base of Knox group.

(1957, p. 407-408) has shown that faults in the foothills belt of the Rocky Mountains in Alberta broke and moved in a manner similar to the examples described here. There are also indications (King and Ferguson, 1960) that the low-angle thrust faults of northeastern Tennessee, originally supposed to have moved on regularly-dipping shear planes (Keith, 1907, fig. 3, p. 9) (fig. 3A), actually moved for long distances over incompetent layers in Cambrian rocks (in the middle of the Unicoi formation, in the Rome formation, and near the base of the Knox group, which are marked 1, 2, and 3 on fig. 3B), and passed more steeply from one to the next through the intervening strata.

If the principles developed from the examples cited are applicable elsewhere, some generalizations regarding the anatomy of low-angle thrust faults can be made, although with awareness that in so varied a group of structures these may not be applicable to all.

(1) Low-angle thrust faults did not originate from low-dipping shear planes which cut indiscriminately through heterogeneous rocks. They followed zones of weakness in the incompetent strata, and shifted abruptly from one zone to the next along diagonal shears in the intervening more competent strata.

(2) The low-angle thrust faults cited, and probably many others, did not originate late in the orogenic cycle in strata already deformed, but earlier, when the strata were little deformed. The thrust sheets moved as broad plates and were not much folded, except where they were warped over the flats and pitches of the fault surface beneath. The overridden rocks were more folded, but probably mainly as a result of frictional drag of the thrust sheet above. Part of the forward movement of the overriding rocks probably resulted from shortening of the overridden rocks by this folding, although doubt has been expressed that it could account for the whole movement (Longwell, 1945, p. 424).

(3) One may suggest, further, that if a low-angle thrust fault developed in deformed rocks of varying competence, its initial rupture could not have been a smoothly-dipping shear plane. Rupture would have been along bodies of incompetent rock, or along zones of weakness already existing. As a simple example, if a thick unit of competent rocks were downfolded into a syncline, the base of the unit in the syncline would not be truncated by the fault, but it would follow the next layer of downfolded incompetent rocks beneath. The resulting fault surface would be complex and its form would be unpredictable; it probably would not have the form depicted in the published sections copied in figure 1.

(4) Several decades ago, it was thought that thrust faults characteristically bring older rocks over younger, and various complex explanations were invoked to explain the exceptional thrust faults in which younger rocks were known to have been brought over older (Billings, 1933, p. 144-153). Thrusting of older rocks over younger is characteristic of most of the exposed thrust faults in the Appalachian and Rocky Mountain regions, but in recent decades many thrust faults along which younger rocks override older have been discovered in the eastern part of the Great Basin (see, for example, Drewes, 1958, p. 228-

230). Along the Pine Mountain and Johnnie-Wheeler Pass faults, older rocks are thrust over younger along the pitches, younger rocks over older along the flats. Both relations are thus normal and expectable along low-angle thrust faults, but occur in different parts of the same fault.

(5) Low-angle thrust faults have finite breadth, and end both rearward and forward. Rearward ends of low-angle thrust faults are seldom available for examination, even in such ideal examples as the Pine Mountain and Johnnie-Wheeler Pass faults, but Nolan's suggestion is plausible that the faults die out rearward by decrease in contrasts between the deformation of the overriding and overridden blocks. Forward, a low-angle thrust fault ascends on a succession of pitches, and the highest pitch will bring it to the surface, over which the forward edge of its thrust sheet will advance along an "erosion thrust" (Hayes, 1891, p. 149-150). These forward edges have long since been carried away by erosion in the Appalachian and Rocky Mountain regions, but they seem to be preserved along many of the thrust faults of the southern Great Basin. This is implied by stratigraphic evidence—formations beneath the thrust plane only a little older than the assumed date of thrusting. It is indicated more clearly by associated deposits—sheared gravels beneath the thrust sheet which probably lay on the ground over which it advanced (Longwell, 1926, p. 570), and sedimentary breccias in front of or partly overridden by the thrust sheet, which were derived from its advancing face (Longwell, 1949, p. 933-935; Ferguson and Muller, 1949, p. 31; Hewett, 1956, p. 99).

(6) Low-angle thrust faults likewise have finite length. A feature of the Pine Mountain fault not hitherto mentioned in this paper is that it has a length of 125 miles, and terminates at the transverse Russell Fork and Jacksboro faults, along which there have been strike-slip movements (Wentworth, 1921, p. 362-368). The Pine Mountain thrust sheet ends against these transverse faults, and the strata beyond them are relatively undisturbed.[2] Even within the thrust sheet itself, as shown by Rich (1934, p. 1591-1594), minor structures indicate zones of potential rupture. Thrust sheets have limits of strength, depending on the materials which compose them, beyond which they cannot move as a single mass. Low-angle thrust faults must end latterally, either along transverse faults with strike-slip displacement, or by loss of displacement along the thrust itself.

The principle is obvious, yet has many times been slighted. Proved low-angle thrust faults in the Appalachian and Cordilleran regions have been extrapolated for scores or hundreds of miles by ingenious linking of possibly unrelated structures, to produce master faults that extend from Pennsylvania to the Alabama coastal plain, or from southeastern Idaho to southern Nevada. Much effort has been expended in discovering northward and southward extensions of the Lewis fault of the Rocky Mountains, grandly displayed as a low-angle thrust near the International Boundary. Whether it maintains this

[2] As shown by Young (1957, p. 2571) some differential movement between the beds has occurred northeast of the Russell Fork fault. Stearns and Wilson (Stearns, 1955; Wilson and Stearns, 1958) have demonstrated even more extensive differential movement and minor thrusting southwest of the Jacksboro fault. These structures should be considered in any detailed treatment of the Pine Mountain fault, but they do not invalidate the principles here stated, and for present purposes can be disregarded.

character far along the strike is doubtful; evidently it deteriorates on either side into a high-angle thrust fault of moderate displacement. In Alberta, the northward extension of the Lewis fault is known to be a high-angle thrust fault behind the McConnell fault, or frontal low-angle thrust fault in the latitude of Calgary (North and Henderson, 1954, map 1).

HABITAT OF LOW-ANGLE THRUST FAULTS

Having discussed details of the anatomy of low-angle thrust faults, we can turn to their larger relations in space, and revert to our earlier question of their habitat, or the environments which were favorable for their development.

The low-angle thrust faults which have been discussed are in stratified rocks of varying degrees of competence. Rocks of this kind, composed of alternating units of carbonate rocks, sandstone, and shales, are characteristic of miogeosynclinal sequences, and low-angle thrust faults are correspondingly abundant in the miogeosynclinal realm. By contrast, rocks laid down in eugeosynclinal areas are slates, graywackes, and volcanics of great thickness, which are less strongly differentiated into competent and incompetent units. Moreover, deformation in the eugeosynclinal realm was mainly accomplished under conditions of so much depth, pressure, and increased temperature that development of the kinds of rupture along which low-angle thrust faults could move was inhibited.

An interesting question is whether there is a casual relation between low-angle thrust faults and the margins of sedimentary basins, or the boundaries between one sedimentary basin and another. Some low-angle thrust faults have transported bodies of older sedimentary rocks over [illegible] of radically younger sedimentary rocks, as along the Lewis fault in Montana (Willis, 1902) and the Great Smoky fault in Tennessee (Neuman, 1951), and these older and younger sedimentary bodies were probably not originally in sequence. Other low-angle thrust faults have transported bodies of sedimentary rocks over other bodies of approximately the same age span, but which are of markedly different facies, as along the Taconic fault of the northern Appalachians (Prindle and Knopf, 1932, p. 293-294), and along the Roberts Mountains fault in north-central Nevada (Roberts and others, 1958, p. 2816-2891), indicating a tectonic superposition of rocks that originally had been laid down contemporaneously in different realms.

Similar extensive transport along low-angle thrust faults has been inferred where contrasting rock sequences of about the same age are juxtaposed along faults, as in the Ouachita area of southeastern Oklahoma (Dake, 1921, p. 55). However, lithologic changes from one sedimentary sequence to another of the same age may be abrupt, as along the margins of the little-disturbed Delaware basin in west Texas (King, 1942, p. 617-623). If faults had developed along the margins of such a basin, markedly different sequences of about the same age would have been juxtaposed, but the contrasts between the sequences would actually be more stratigraphic than tectonic. Juxtaposition alone is thus not a conclusive indication of extensive transport along low-angle thrust faults; this requires corroboration from other structural evidence.

Thus some of the inferred relations between low-angle thrust faults and

the margins of sedimentary basins may be more apparent than real. Location of the faults may be accidental, and telescoping along the faults may simply accentuate large stratigraphic variations already existing. The observed occurrences are sufficiently common, however, to suggest that the lay of the strata along the margins of a sedimentary basin may have guided the initial ruptures along which the thrusts subsequently moved. An interesting analysis of the stratigraphic relations of a succession of thrusts along the margin of a sedimentary basin in western Wyoming has been made by Rubey and Hubbert (1959, p. 189-191). Similar analyses are no doubt available for faulted basin margins in other regions. Other analyses are desirable, although the necessary data are not everywhere readily obtainable, because of extensive loss or concealment of the transitional strata as a result of the thrusting.

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ANALYSIS OF SOME RECENT GEOSYNCLINAL THEORY

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ABSTRACT. "Geosyncline" was invented by Dana in 1873. It was defined as "a down-bending of the crust"; the context supplied the connotation that a mountain chain would eventually arise from sediments accumulated in such a down-bending. However, "geosyncline" has not yet been adequately defined.

Stille's 1940 classification of geosynclines was introduced into America by Kay in 1942. Stille's two principal classes are based upon capacity to be folded. Highly foldable geosynclines are orthogeosynclines; weakly foldable, parageosynclines. Orthogeosynclines include eugeosynclines and miogeosynclines, distinguished by position, time of folding, and magmatism. Eugeosynclines supposedly determine the sites of autochthonous palingenic bathylithic magma. However, bathyliths occur also in so-called miogeosynclines, which necessitates the auxiliary hypothesis that magmas can travel laterally from their eugeosynclinal place of origin to a miogeosynclinal environment. Recent absolute age determinations on bathyliths render this hypothesis untenable.

Lawson maintained that geosynclines do not collapse until they have accumulated a critical thickness—approximately 40,000 feet. Nevertheless, geosynclines have been shown to fail under loads of 10,000 feet or less. The *maximum* thickness of accumulation appears to be 40,000 feet; consequently this limit suggests that stratigraphic thicknesses estimated to exceed 40,000 feet require reduction.

In 1944 Stille introduced the concept of "regeneration," meaning thereby restoration of the orthogeosynclinal state in ancient fold-belts. This idea has stimulated, among other things, active interest in the fate of ore deposits in regenerated orthogeosynclines. Critical evidence of the effects of regeneration on the dispersal of pre-existent ores is given by the isotopic composition of leads in galenas formed during successive orogenies.

ORIGIN OF THE TERM GEOSYNCLINE

The term geosyncline, as Goguel remarked in 1952, has had a brilliant career ever since its invention by Dana in 1873. In 1940, several kinds of geosynclines were defined and named by Stille, who also devised a classification for them. Shortly afterward, many more kinds of geosynclines were distinguished and named by Kay. It is proposed here to analyze the content of the more widely used of these terms and to show their impact on other major problems of geology, such as the origin of primary and palingenic magmas and the origin and classification of ore deposits.

The *idea* that a trough containing a thick accumulation of sediments was destined to become the site of a future mountain chain was originally formulated by James Hall, State Paleontologist of New York, in his presidential address to the American Association for the Advancement of Science in 1857, and was presented at greater length in 1859 in volume 3 of the New York Geological Survey (Hall, 1859).

The *name* was created by Dana in 1873 (Dana, 1873, p. 430) in the form of "geosynclinal". Dana, however, gave the term no comprehensive definition. He defined geosynclinal briefly in a footnote on page 430: "from the Greek ge, the *Earth*, and *synclinal*, it being a bend in the Earth's crust". The definition was given incidentally while Dana was explaining how the Allegheny Range was formed:

> The making of the Allegheny Range was carried forward at first through a long-continued subsidence—a *geosynclinal* (not a true geosynclinal since the rocks of the bending crust may have had in them many true or simple synclinals as well as anticlinals), and a consequent accumulation of sediments, which occupied the whole of Paleozoic time.

The mountain range formed from the folding of such a geosynclinal accumula-

tion was called a synclinorium, from *synclinal* and the Greek *oros*, mountain. Etymologically better, it should have been termed a "geosynclinorium". Both Dana and Hall before him were mainly concerned with the problem of mountain-making. Hall, in finally publishing his presidential address in 1882 (Hall, 1882, p. 68) expressly affirmed that the address of 1857 was intended to show "that mountain ranges were coincident with lines of great sedimentary accumulation". Thereby the idea became implanted in geology that a geosyncline and its accumulation of sediments predestines a future fold chain. This idea can fittingly be called, with perfect etymology, the "geosynclinorial theory" of Hall and Dana.

RELATION OF GEOSYNCLINES TO OROGENY

In the theory of mountain-making based upon the geosyncline, Dana recognized two stages: (1) in which a "preparatory" geosyncline was gradually deepening and concurrently accumulated a great thickness of sediments, the building material of the future mountain range; and (2) the mountain-making events—the folding and faulting of the accumulated strata. In the final edition of his Manual of Geology, Dana (1895, p. 357) wrote:

> The knowledge of the Appalachian facts led Professor James Hall to suggest in 1859 that a similar trough of deposition preceded the up-turning in all cases of mountain-making. It was the first statement of this grand principle in orography.

Thus by 1895 the idea had become established that all geosynclines that had been the loci of heavy sedimentation are "preparatory" geosynclines, and that such geosynclines are consequently the sites from which future mountain chains will arise. A great principle of orogenesis had indeed been stated, but no generally satisfactory definition of geosyncline had then been formulated, nor has one yet been produced.

MEANING OF THE TERM GEOSYNCLINE

Definitions range widely. Stille (1924, p. 6; 1940, p. 164) says that a geosyncline, in the broadest usage, is a secularly sinking basin of sedimentation. He adds that the formation of a fold-range from a geosynclinal accumulation is not necessary to the concept "geosyncline", thereby departing from the original ideas of Hall and Dana. On the other hand, Termier and Termier (1956, p. 2) in a treatise of world-wide scope define a geosyncline thus: it is a subsiding zone that, as a rule, has received a thick sedimentary accumulation, has then been subjected to regional metamorphism and to orogenic phases, and finally is transformed into a chain of mountains.

In 1936 "pliomagmatic" and "miomagmatic" were proposed by Stille (1936) to characterize certain types of geosynclines. Regions such as the Nevadan realm of North America that are characterized by strong volcanism during sedimentary accumulation and by synorogenic plutonism during the subsequent deformation were called "pliomagmatic". These terms, so the author insisted, are to be applied only to geosynclines that had been capable of being closely folded or, to use his own words, they are to be used to characterize the "alpinotype" condition of a region and not its subsequent "germanotype" condition after it had become consolidated and was no longer capable of being folded and closely appressed.

In 1940 "pliomagmatic geosyncline" was shortened by Stille (1940) to "eugeosyncline" and "miomagmatic geosyncline" to "miogeosyncline". The content of the terms was also amplified. The terms were defined in a book entitled "Einführung in den Bau Amerikas" published in 1940 (Stille, 1940). The volume contains also a valuable glossary of terms. World War II had already begun, and so far as known to Dr. Stille[1], only four copies of the volume reached the United States of America. Shortly afterwards the whole stock at Borntraeger was destroyed during an air raid on Berlin.

STILLE'S CLASSIFICATION OF GEOSYNCLINES

Stille's terms first appeared in American literature in a paper by Marshall Kay (1942, p. 1642). In a footnote to that paper on p. 1642, the following clasification of geosynclines was presented, as given by Stille in a letter written to Kay in German under date of May 24, 1941. As translated here by me the Stille classification reads:

I. Orthogeosynclines (alpinotype geosynclines) comprising
   a. Eugeosynclinal zones (pliomagmatic mother zones of the internides)[2],
   b. Miogeosynclinal zones (miomagmatic mother zones of the externides)[3].

II. Parageosynclines (germanotype geosynclines upon already consolidated basements).

To which Kay added: "the terms eugeosyncline and miogeosyncline have been defined recently (Stille, 1940[4], p. 15)".

The term "parageosyncline" had previously been used by Schuchert (1923, p. 199) in another sense, but it had not come into use. The choice of the term parageosyncline by Stille was hardly felicitous, since "para" is used in so many senses in geologic terminology. As in Stille's definition parageosyncline is defined as being situated upon an already consolidated basement, in other words, upon the craton, "intracratonic or supracratonic geosyncline" would have been preferable.

The terms formulated by Stille in 1940 were adopted by Kay in 1942. To them Kay added other terms to specifically designate the geosynclines of many kinds that he proposed to distinguish. Although no one has yet been able to give a generally acceptable definition of "geosyncline", we now have thirty or more variants of the unsatisfactorily defined concept.

As commonly used in America, the terms eugeosyncline and miogeosyncline are purely descriptive. Eugeosyncline is employed to designate a thick stratigraphic section, i.e., measurable in thousands of feet, that contains abundant contemporaneous volcanic rocks. Miogeosyncline is used to signify a thick stratigraphic section devoid or nearly devoid of contemporaneous volcanic

[1] Letter to the writer (A.K.) under date of June 2, 1959.

[2] Internides: The inner and therefore the older zones of alpinotype fold chains.

[3] Externides: The outer and therefore the younger zones of alpinotype fold chains. In geosynclinal chains on the borders of continents the externides are on the continental side.

[4] Originally cited in error by Kay as 1941. The definitions referred to were not given.

rocks. Paleogeographic and stratigraphic evidence of alleged geosynclines are generally sketchy or are not given.

In a recent publication that is easily available Stille (1958, p. 284) again defined "orthogeosynclines" (an overall category that, as already mentioned, comprises both eugeosynclines and miogeosynclines) as those mobile zones of the Earth's crust in which have been concentrated the alpinotype foldings and the shortening of the crust thereby produced. While "eugeosyncline" and "miogeosyncline" were becoming widely used in America, the author of the terms came to use them only sparingly or not at all. In a masterly synthesis of the late Paleozoic Variscan orogeny as it affected the continent of Europe—a monograph of 138 pages,—Stille (1951) employs neither term; instead "inner Variscan (Innenvariszikum)" and "outer Variscan (Aussenvariszikum)" are used to distinguish these portions of the geosynclinal system "on the basis of time of folding, intensity of folding, and magmatism".[5]

Many authors in employing the terms eugeosyncline and miogeosyncline casually explain that miogeosyncline means "less of a geosyncline", but they do not explain what is meant by "less of a geosyncline". In the words of Stille (1940, p. 656), who created the term, miogeosynclinal zones are

> more peripheral, generally miomagmatic or amagmatic elongate belts of a geosynclinal system, which manifest a certain 'lesser geosynclinality' in that they commonly become folded only after their neighboring eugeosynclinal zones have become folded.

The power to make such a distinction obviously requires a correct and comprehensive knowledge of the geologic history of the geosynclinal system in question.

MAGMATISM AND OROGENY IN RELATION TO GEOSYNCLINES

Initial magmatism, meaning thereby the eruption of lavas and the intrusion of related dikes, sills, and stocks during the accumulation of a geosynclinal prism, is regarded as characteristic of an orthogeosynclinal system. This igneous activity is specifically termed by Stille "initial magmatism". Such magmatism does not characterize all zones of an orthogeosyncline, however. Typically, initial magmatism is absent from the outer zones of an orthogeosynclinal system; that is, it is absent in those zones that during the collapse of the geosyncline become folded only during the later phases of a polyphase orogeny. In this usage the term "orthogeosyncline" includes; (1) the idea of "full geosynclinal (eugeosynclinal) state", which is characterized by the attributes of high geosynclinality, one of which is manifested most obviously by initial magmatism; and (2) a "lesser geosynclinal" state, in which initial magmatism has not participated. According to Stille, the eugeosynclinal zones have in general been folded first and have therefore been folded far more strongly than those portions of the orthogeosynclinal system that were folded later. The eugeosynclinal zones, which are the loci of initial magmatism, are said to have been as a rule the principal loci of synorogenic plutonic intrusions, but as shown later, this generalization needs considerable qualification. The eugeosynclinal zones are thus likely to be doubly pliomagmatic in contrast to the miomagmatic or amagmatic outer zones. The reason for this, according to

[5] Letter to the writer, June 2, 1959.

Stille, is that the closely folded belt produced during the deformation of the geosyncline sinks so deeply that the lower portion fuses to form palingenic sialic magma. After the upper portion of the folded belt has become rigidified—when it has attained a quasi-cratonic state—it may become the locus of "subsequent magmatism", marked by eruption of consanguineous sialic lavas and the intrusion of subvolcanic plutons. Therefore the original geosynclinal zone can be characterized as pliomagmatic on a third count.

The term eugeosyncline therefore embodies three features: (1) the eugeosyncline is that portion of a geosynclinal system which was the first to be folded; (2) it comprises the portion of higher mobility, i.e., the capacity to be folded and closely appressed; and (3) it is the locus of abundant igneous activity, both during the preparatory stage of the geosyncline and its subsequent conversion into a folded tract, or orogen as it was named by Kober (1921, p. 21). The orogen, as Kober ventured to emphasize, is from one-third to one-half narrower than the original geosyncline. "Orogen" has at least this advantage over "geosyncline", it is one stage nearer factual reality than is "geosyncline", in which the contained load is inferred to have been the necessary condition determining that a fold-chain has arisen from it.

Special insight, often of a predictive nature, has become linked to the term eugeosyncline. As early as 1942 Kay announced that the eugeosynclines become the loci of bathylithic intrusion of granites. "Why", asks King (1959, p. 147) "are granitic rocks characteristically emplaced only in the former geosynclinal areas?" The idea underlying this question is a variant of an earlier generalization. Said Browne (1932, p. 135) "Wherever a bathylith outcrops we may infer the former presence, on or over that spot, of a series of geosynclinal sediments, terrestrial or marine, during whose folding and elevation the intrusion of the bathylith took place". As of 1959 we can rephrase Browne's generalization thus: "Wherever a bathylith outcrops we may infer the former presence, on or over that spot, of a series of *eugeosynclinal* rocks."

On several counts, however, even the revised generalization is too inclusive. Bathyliths occur also in non-orogenic regions, as exemplified by the Rand granite of the Bushveld Igneous Complex, the granite comprising a mass of 5000 square miles in area; and by the Rapakivi mass near Viborg, Finland, 6400 square miles in area. Furthermore, bathyliths do occur in what have been called miogeosynclines, such as the Idaho bathylith of 18,000 square miles, the Boulder bathylith of Montana of 1200 square miles, the bathyliths of the Philipsburg district, Montana (Emmons and Calkins, 1913), and the large granitic stocks of almost bathylithic dimensions in the Wasatch Range, stocks which penetrate the thick amagmatic series of sedimentary rocks that accumulated in the Wasatch geosyncline.

The occurrence of large bathyliths in the Rocky Mountain region was regarded as anomalous by Stille. The folding was weak; consequently downfolding plus isostatic sinking was inferred to have been not deep enough to cause fusion of the geosynclinal prism. Because the magma could therefore not have been formed subjacently, it was considered to have moved in laterally from a supposed locus of generation farther west. The idea of eastward migration of magma had previously been advanced by Lindgren on other grounds.

Stille calls bathylithic intrusion of this kind "allochthonous synorogenic plutonism fed by the lateral migration of sialic magma." In an account of the Boulder bathylith (Knopf, 1957, p. 88) I presented these ideas without comment, believing them to be highly interesting but unsupported by compelling evidence. Since that time reliable age datings by the K-Ar method have become available. They demonstrate clearly that eastward migration of magma in time is *not* supported by the evidence; for example, the granites in the Coast Ranges of California are 80 to 90 m.y. old; the granites of the foothill region of the Sierra Nevada are 140 m.y. old; the granites of the higher Sierra Nevada are 80 to 90 m.y. old; and the Boulder bathylith, far to the east in Montana, is also 80 m.y. old.

CRITIQUES OF THE GEOSYNCLINAL DOCTRINE

Some recent critiques of the geosynclinal doctrine should be noted. Goguel, (1952) after commenting that Dana's term "geosynclinal" has had a brilliant, not to say dazzling, (*èclatant*) career, finds that it is impossible to give a precise meaning to the content of that concept. The essential idea implied by the term is that the sediments in a geosyncline are very much thicker in the folded zone than elsewhere. "But", added Goguel, "the other characteristics of sedimentary accumulations called geosynclinal vary with the author". It takes no worldwide survey of geologic literature to find this statement amply confirmed.

The Appalachian geosyncline, as is well known, is the type geosyncline. However, the great increase in knowledge of the region, including the many absolute age determinations recently made, which fix the dates of metamorphism and bathylithic intrusion beyond argument, show that the geologic history of the region differs greatly from that of the orthodox belief. When the recognition by Kay (1944, p. 462) that "there are geosynclines of several sorts and of differing ages in the Appalachian region", becomes generally accepted a great advance will have been made.

RECOGNITION OF MODERN GEOSYNCLINES

As has often been pointed out, the fundamental principle of orogeny that a geosyncline and its load of sediments predestines a future fold chain would prevent us from recognizing a modern geosyncline that has not yet been folded.

Nevertheless, it is probable that modern geosynclines can be recognized, but no criteria have yet been established by which to determine whether or not such modern geosynclines are "preparatory geosynclines" out of which fold-chains will eventually arise.

Kossmat (1936) thought that the basin of the Adriatic Sea is a typical geosyncline, still sinking. Kuenen (1935) considered the Timor trough and its continuation as a geosyncline to be a "modern geosyncline" in the act of formation.

The Gulf Coast region of Louisiana and Texas is considered by many geologists to be a modern geosyncline, despite its failure to agree with the orthodox conception of what the type geosyncline—the Appalachian—was like. To Bucher (1951) the so-called Gulf Coast geosyncline is a particularly fine representative of a modern geosyncline; to Lawson (1942) it has the most

significant feature of a geosyncline, namely the heavy load of sediments acquired *pari passu* with subsidence, i. e., the belt of excessive sedimentation which as emphasized by Hall, the founder of the geosynclinal theory, was the factor determining the site of the future mountain range.

According to Crouch (1959, p. 1290), however, the Gulf Coast "geosyncline" is in the unfortunate state that "none of the requirements for a geosyncline can be substantiated at this time; that is, no regional reversal of dip and no thinning of beds can be demonstrated". Finally, we may cite King (1959, p. 82), who believes that the Gulf Coast deposits do not differ very fundamentally from those of the Appalachian geosyncline, and that the differences "are mainly the result of differences between the internal geography of the continent in Paleozoic and Cenozoic time".

THE CRITICAL LIMIT OF THE FILLING OF A GEOSYNCLINE

In a series of papers beginning in 1927, Lawson (1927) advanced the idea that the sedimentary trough collapses only after the filling has reached a certain critical limit, which he placed at approximately 40,000 feet. Earlier however, Schuchert (1923, p. 208), having inductively examined the recorded field evidence of the failure of geosynclines, concluded that the thickness of the sediments accumulated in a geosyncline appears to have nothing to do with the time when the sediments were folded; in some folded belts the sediments appear to have been less than 10,000 feet thick.

Goguel in his *Traité de Tectonique* (1952, p. 23) goes even further than Schuchert, pointing out that we know of basins containing exceptionally thick sediments that have not been folded, as well as folded zones in which the sedimentary series are of feeble thickness.

Of still greater importance, because supported by examples, is the evidence cited by Lees (1952, p. 4). He finds that deep exploratory borings have shown that, contrary to expectation, many sequences of sedimentary rocks in flat-lying or very gently folded foreland areas are even thicker than their equivalents in the strongly folded or thrust mountain belt.

According to a remarkable summary account by Browne (1947, p. 635), the Tasman geosyncline of eastern Australia, because of its length (2000 miles), its duration in time (the whole of the Paleozoic), and its eventful history, "must surely rank as one of the greatest of its kind that the world has ever seen". The geosyncline was subjected in its various parts to five major orogenies, all of which were accompanied by granitic intrusion. The fourth of these orogenies, the Kanimblan, occurred near the end of Mississippian time. It is considered to be the greatest in the history of the Tasman geosyncline. The folding extended 2000 miles in length and affected the paralic zone and some of the marine zone of the geosyncline over a total width of 600 miles. The Kanimblan folding, although as a rule comparatively gentle, was followed by bathylithic intrusion in many of the areas affected (David and Browne, 1950, p. 319). Yet the thickness of the Devonian and Carboniferous beds that were folded is reported to be less than 10,000 feet.

In view of these statements by Schuchert, Goguel, Lees, David and Browne, it appears probable that Lawson's term "critical limit" must be

changed to mean not that folding failed to ensue until the sedimentary filling of the geosyncline reached 40,000 feet in thickness, but that no sequence (sedimentary or volcanic) can exceed about 40,000 feet in thickness without being folded. The following geosynclines are reported to have contained accumulations aggregating 40,000 feet before being folded: Appalachian; Llanorian (35,000 feet); Eastern Venezuelan; and Wasatch, Utah. On the other hand, geosynclinal sections that are reported to be much thicker than 40,000 feet prove on careful restudy to be generally, if not invariably, far thinner than originally announced.

RESTORATION OF THE ORTHOGEOSYNCLINAL CONDITION

An important new concept was introduced by Stille (1944, p. 195) into geosynclinal theory. An account of this idea was later (Stille, 1955) given in English. The concept is called "regeneration" or the restoration of the geosynclinal state. During regeneration a folded tract recovers the orthogeosynclinality that had been lost by the operation of preceding orogenic processes. When such regeneration takes place it affects not only the structures in the crust, but even more the magmatic relationships. An unexpected result of the introduction of this concept into geosynclinal theory has been the powerful stimulus it has exerted on theories of the origin of ore deposits.

In 1921 de Launay (1921, p. 333) pointed out that the fold chains of France are built upon the ruins of older fold-chains. Said he: "Nature has acted here just as does man who forever continues to build upon the preceding ruins". The Alps stand upon the ruins of a Variscan chain, and the Variscan chain stands upon the ruins of a still older chain, probably of Assyntian construction at the end of the Precambrian, and possibly on a still older one, according to Stille (1951).

The earliest recognition of this general idea, so far as known to me, is by P. Termier (1903). He pointed out that the Alpine geosyncline was established upon a zone that already had been folded before the deposition of the Carboniferous coal measures, and that the axis of the geosyncline is essentially parallel to the Variscan folds. However, in the Mont Blanc region, where at least four phases of folding have occurred, the axes of the youngest folding of the mid-Tertiary Alpine orogeny are not parallel to those of the older folding (Oulianoff, 1934, p. 125). Such axial divergence between the Alpine and older folding is evidence to Stille (1951, p. 41) that the Alps arose out of a regenerated orthogeosyncline and not from a residual portion of a Variscan or older geosyncline.

According to the French view, a fold-chain in the course of time is reduced to a peneplain, from which by subsidence, sedimentation, and folding a new chain will arise. A geotectonic cycle thus begins with a peneplain and ends with a peneplain. De Launay's generalization has been amplified so as to say that most mountain chains are built upon the ruins of older chains. In that generalization doubtless resides the explanation of why "the basement of all geosynclines is found to be a continental area" (Kuenen, 1950, p. 145). Earlier, Kuenen (1935) already had announced that geosynclinal troughs invariably develop on continental areas underlain by a thick sialic crust. Logically then,

the conclusion follows that throughout geologic time the development of such troughs was a consequence of restoration or recovery of orthogeosynclinality.

Exploration of our oldest Precambrian terranes may possibly in time supply us with evidence that there have been geosynclinal troughs which had not been developed from continental material. These geosynclines will probably fall into the class of ensimatic geosynclines, which in theory contain an accumulation consisting of effusive rocks that were poured out on a sima floor (Wells, 1949).

The Alps have long been recognized to stand on the ruins of older mountains: on Variscan (Hercynian) of late Paleozoic age and in part on still older remnants, probably of late Precambrian age (Assyntian). The region affected by the Tertiary Alpine orogeny is not a residual portion of the Variscan geosyncline or of a pre-Variscan geosyncline but is a regenerated orthogeosyncline. The restoration of foldability to the extreme degree shown by Alpine folding is unsurpassed in any of the world's orogenic belts.

An example recently recognized in Australia is described by Hills, (Hills, (1956). The Adelaidean geosyncline, of late Precambrian and early Paleozoic infilling, is a geosyncline that lies between blocks of cratonic basement, which had previously been strongly and universally folded.

REGENERATION IN ITS APPLICABILITY TO ORE DEPOSITS

The concept of the recovery of geosynclinality has led Schneiderhöhn (1952) to formulate a new classification of ore deposits, which is based on (1) the older magmatologic considerations, (2) the newer geotectonic ideas of Stille, and (3) the role of metamorphism as developed by Clar, Angel, and Friedrich. In this classification:

Class I comprises essentially the deposits formed in a primary orogen by fluids of magmatic derivation,

Class II regenerated ore deposits, comprise deposits of Class I whose constituents were put in circulation again during geotectonic regeneration of the primary orogen and were deposited in higher and younger parts of the Earth's crust.

These ideas have led to great intellectual activity. Space does not permit further discussion, but I must content myself with a brief mention of an extraordinarily penetrating account of the application of Schneiderhöhn's ideas by Smirnov (1959) to the ore deposits of the Caucasus. In that region tectonic, magmatic, and metallogenic processes took place during the Caledonian, Variscan, and Mesozoic-Cenozoic (Alpide) orogenies. Molybdenum deposits were formed during all three orogenies. At first sight this thrice repeated formation of molybdenum deposits would seem to support Schneiderhöhn's hypothesis. However, the first two metallogenic epochs produced only insignificant molybdenum deposits, but during the Alpide epoch important ore bodies were formed. The conclusion that the not inconsiderable number of Alpidic ore deposits of the Caucasus were formed according to the scheme of regeneration from the very small number of older deposits "is completely impossible". The same rule holds for the other metals. The reserves of lead and zinc in the known

Alpide ore deposits are at least 15 to 20 times larger than those in the Variscan deposits. It is therefore improbable that the lead and zinc in the Alpide deposits were derived from the Variscan deposits. This conclusion of Smirnov is independently confirmed by the differing isotopic composition of the lead in the galenas of the deposits of both ages.

Isotopic Composition of Leads

	Pb^{206}/Pb^{204}	Pb^{207}/Pb^{204}	Pb^{208}/Pb^{204}
Variscan	18.02	15.22	37.31
Alpide	18.60	16.08	39.13

The Variscan lead corresponds well in isotopic composition with that of normal late Paleozoic lead, and the Alpide lead with normal Tertiary lead.

However, many examples in which the isotopic composition of a lead does not correspond with its geologic age have become known. In the late Mesozoic ore deposits of Bisbee, Arizona, the lead has an isotopic composition indicative of Precambrian age (Bain, 1952). Manifestly, the late Mesozoic ore-forming solutions had acquired from somewhere in depth lead that had previously been deposited in Precambrian time, had carried it upward, and had redeposited it in rocks higher in the crust. Many more examples of such "anomalous leads", as they are termed by Russell, et al. (1954, p. 305) could be cited; however, the geologic explanation of their anomalous compositions is still in the realm of the future.

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ORIGIN OF THE PACIFIC ISLAND MOLLUSCAN FAUNA*

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ABSTRACT. It has long been recognized that the marine mollusks of the Pacific islands are related to the fauna of Indonesia, there being few ties between the mollusks of the islands and those of western America. Indonesia has been regarded as the center of dispersal of the "Indo-Pacific fauna", yet prevailing winds, currents and even the major storm tracks trend from the islands toward Indonesia and the west.

Until recently all of the islands of the Pacific Basin were thought to be geologically very young, but data obtained from submarine mapping, dredging, and drilling now indicate that: (1) a shallow water fauna, including mollusks, was present in the area as early as middle Cretaceous; (2) rich Tertiary molluscan faunas (possibly richer than those of today) were widespread both inside and outside the Pacific Basin proper; (3) many, if not all, of the hundreds of atolls that lie between Hawaii and Indonesia stood above the sea (some as high islands) at intervals during the Tertiary; (4) some 50 guyots, now far beneath the sea, projected into shallow water as additional stepping stones for the distribution of marine life, many of them in the broad gap southwest of Hawaii. These discoveries suggest that the Pacific islands once formed a giant archipelago and that the islands could have been the home of many elements of the Indo-Pacific fauna. Faunal migration, favored by winds and currents, was toward Indonesia rather than from it.

With some islands or parts of islands projecting above the sea at all times since the Cretaceous there would be a reasonable explanation for the ancient stocks of land shells and plants found in Hawaii and other volcanic islands. The newly discovered stepping stones appear to be satisfactory replacements for the land bridges once called for by many biologists.

INTRODUCTION

Area

The Pacific Ocean, exclusive of adjacent seas, covers 64 million square miles (Shepard, 1948, p. 281) or about one-third of the earth's surface. Pacific islands, including New Guinea, the large continental islands of Indonesia and the "island continent" of New Zealand cover only about 1,400,000 square miles or less than 3 percent of the ocean area. Many of the islands, including most of the large ones, are concentrated in the southwest quadrant. To the north and east of centrally located Hawaii there are wide areas of unbroken ocean (fig. 1).

The island area, as here described (fig. 2), includes a generous southwest one-quarter of the Pacific basin. It extends from Hawaii to the west, southwest, and south to include almost all of Polynesia, Micronesia and Melanesia. Under existing conditions this area cannot be looked upon as a gigantic archipelago because it includes groups of islands that are separated by hundreds of miles of ocean containing neither islands nor reefs. Such gaps, however, did not exist in past times when hundreds of islands that are now atolls projected above the sea and volcanic mounds (guyots) that are now deeply submerged projected into shallow water to form banks and reefs.

Faunal relations

Intensive studies of the distribution of life in the Pacific date back to Charles Darwin and Alfred Russel Wallace one hundred years ago. Each of these co-discoverers of the theory of natural selection obtained support from

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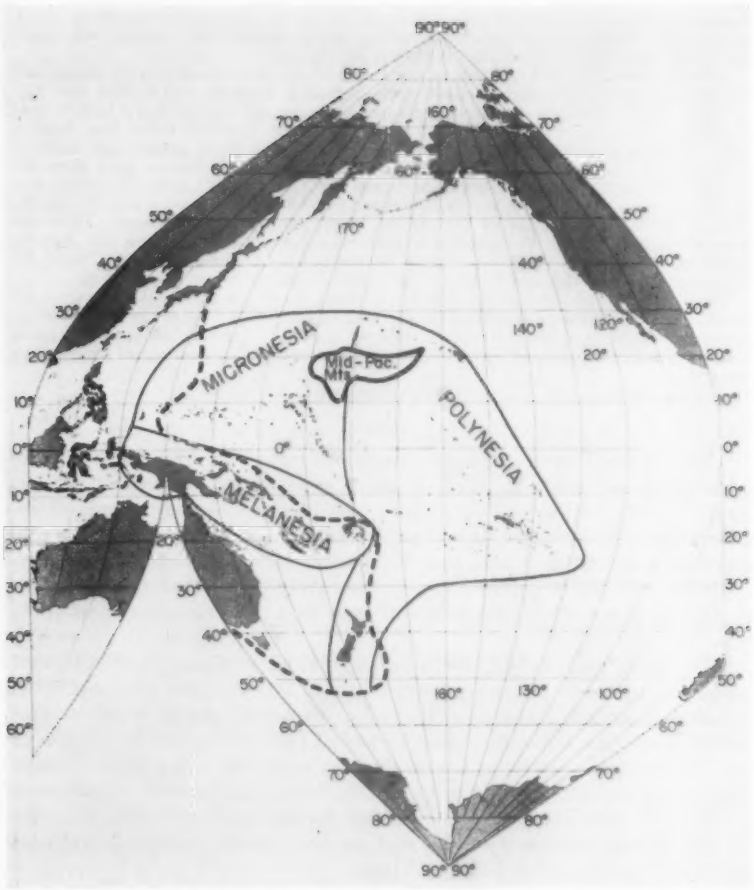

Fig. 1. Map of Pacific showing boundaries of Polynesia, Micronesia, Melanesia; andesite line shown by dashes. Base map from data furnished by Hydrographic Office.

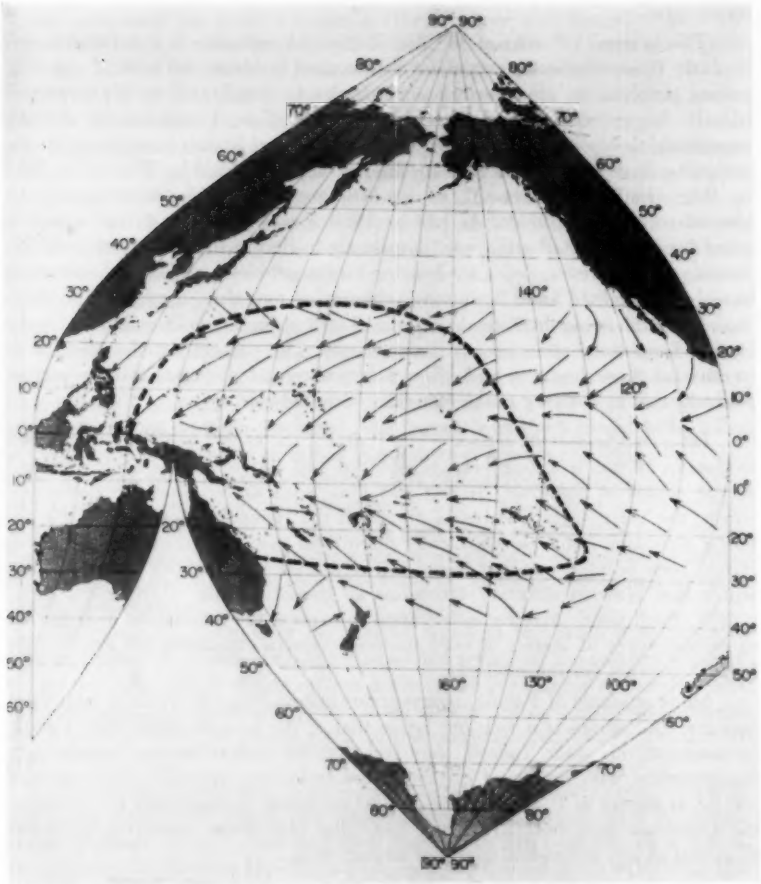

Fig. 2. Prevailing winds in and near island area (dashed line). Figure shows pattern in January; in July there is a shift to the north but no essential changes are involved. Data from Tannehill, 1952, with permission of Princeton University Press.

island studies, Darwin in the Galapagos (1839) and Wallace in the East Indies (1860). Both concerned themselves chiefly with vertebrates but their interests included insects and other invertebrates. The patterns of distribution that they observed were explained by evolutionary theory and strongly supported it. The unique advantages of islands in evolutionary studies were recognized in their classic studies.

The general patterns of the distribution of mollusks in the Pacific, particularly those of the terrestrial forms, aroused attention because of the difficulties involved in transporting such forms to small and widely scattered islands. Suggested dispersal agents have included land connections, drifting vegetation, typhoons and migratory birds. The use of islands as stepping stones, including those now buried beneath the sea, was suggested by Wallace in 1881 (p. 270). In Wallace's time there was little geological evidence to support the idea of submerged islands. As late as 1950 it was pointed out that complete proof for island distribution was "hopelessly buried in the geological past." In recent years, however, new data bearing on the problems of island distribution have been obtained from 3 sources—submarine mapping, dredging and drilling. Much is being learned about the distribution of molluscan life in the Pacific from the Cretaceous to Recent times. The present article attempts to summarize these findings and offers a hypothesis of dispersal for the marine mollusks that is contrary to the generally accepted conception.

Indo-Pacific marine fauna.—The Indo-Pacific aspect of the Pacific island molluscan faunas has long been recognized and Indonesia has been regarded as the center of dispersal. It has also long been known that the prevailing winds, the main surface currents and even the typhoon tracks are in the opposite direction, but these apparently discordant facts have often been ignored. It has been pointed out that there are high level winds above the prevailing winds that blow in different directions at greater speeds (Darlington, 1957, p. 20). Such winds may be called upon to explain the anomalous distribution of certain types of insects or even minute terrestrial mollsuks but are of little help as regards marine mollusks.

Many students of Pacific invertebrates believe that the fauna of Indonesia and the Philippines is a rich one while that of the islands of the open Pacific is comparatively poor. Indeed, they feel that the faunas become increasingly impoverished with increased distance from Indonesia. Actually, the case may not be as strong as it appears for it must be borne in mind that the fauna of the Indonesia area, both living and fossil, has been more intensively collected than that of any individual island or island group.

Consider, for example, the reef-building corals. A map showing the distribution of genera in 39 localities was published by Wells (1954, pl. 186). More than 50 genera have been reported from the much studied Great Barrier Reef, parts of Indonesia (Celebes) and nearby Palau. To the north and east the numbers of genera drop, Hawaii having only 15. It is notable, however, that a prominent bulge in the isopangeneric lines had to be drawn to take in the Marshall Islands where 52 genera were found, mostly from intensively collected Bikini. If all islands from Indonesia to Hawaii were as closely collected as was

Bikini the pattern of distribution as shown by numbers of genera might be greatly altered.

The marine molluscan fauna of the Indo-Pacific is known to be large. Ekman (1953, p. 13) estimated it at 6,000 species. The known faunas of individual island groups are appreciably smaller. A total of about 1,000 shelled forms were listed for French Oceania (Dautzenberg and Bouge, 1938). The known Hawaiian fauna is probably close to 1,500 species. Solem thinks it probable that the list for the New Hebrides will eventually be well over 2,000 species (1959, p. 267). Actually the fauna of most groups is incompletely known and this is particularly true of the micro-mollusca.

Information about the recently discovered shallow water Cretaceous fauna dredged from 2 guyots in the now submerged Mid-Pacific Mountains between Hawaii and the Marshall Islands (fig. 1) is given by Hamilton. The assemblage is made up of reef corals, stromatoporoids, rudistids and other mollusks (Hamilton, 1956). During the Eocene, shallow water limestones containing numerous benthonic larger Foraminifera were laid down on widely separated islands—in Tonga, the Marshalls, the Marianas and Palau. Mollsuks also occur in many of these limestones but well preserved material has not yet been found. In the Miocene sediments, the molluscan records are much more complete and preservation is excellent. Abundant and varied molluscan assemblages have been collected from the Marshalls, Fiji, the Marianas and Palau, though most of these have not yet been described. The faunas from the lagoonal beds drilled in the Marshalls appear to be richer in numbers of species and individuals than those dredged from the same lagoons today. The lagoonal faunas, both living and fossil, are dominated by the gastropods but in other areas where volcanic muds are found there is a rich fauna of pelecypods as well. It is possible that the molluscan fauna of the existing sea in many parts of the island area is poorer than it was during parts of the Miocene, although this cannot as yet be clearly demonstrated.

The observed richness of the marine fauna in Indonesia may be partly due, as Fenner A. Chace, Jr. suggested in his review of the present paper, to the accumulation there of species that arose in the island area and drifted west and southwest. All those that found a suitable niche in hospitable Indonesia would survive after the parent stock in the ever-changing islands became extinct.

Terrestrial fauna and flora.—Interpretations involving the terrestrial fauna and flora of the islands differ greatly from those dealing with the marine fauna. Students of the varied land shell faunas that are found on all of the existing high islands from Hawaii to Indonesia recognized at an early date that these faunas appear to have been derived from ancient stocks. Pilsbry pointed out long ago that the faunas, though containing many endemic species, were nearly homogeneous over wide areas and contained no admixture of the great series of modern families that characterize Tertiary and Recent continental faunas. Pilsbry felt it necessary to postulate a late Paleozoic or early Mesozoic continent upon which he superposed the present island masses of volcanic and coral rock (Pilsbry, 1900, p. 581). Adamson in a review dealing primarily with the terrestrial fauna of the Marquesas agreed with Pilsbry's appraisal of

ancient lineage, stating pointedly that no modern family of land snails reached the central Pacific until brought by man (1939, p. 15). The high degree of endemism in the Marquesas fauna suggested that the islands had been an isolated archipelago since early Tertiary, if not earlier (p. 75). Adamson favored wind distribution rather than land connections.

Paleontological evidence bearing on the history of Pacific land shells is meager but suggestive. Two species of typical high island shells have been described from the Miocene of the Marshall Island drill holes (Ladd, 1958). One of these species is from the lower Miocene, so it is certain that some land shells were present in the islands in middle Tertiary times. In Indonesia, on the other hand, the richly fossiliferous and much studied Tertiary sections of Java have yielded no land shells below the upper Pliocene (Jutting, 1937, p. 171).

Studies of Pacific island insects have revealed some unusual patterns and have resulted in a variety of interpretations. Some have called for land bridges (Meyrick, E., 1899, p. 132), others for island stepping stones (Zimmerman, 1948, p. 49-52), still others for continental drift (Britton, 1957, p. 1383-1389). Usinger noted that in many groups of insects existing today there appears to be a progressive diminution in the number and variety of forms from the rich epicontinental areas of the southwest Pacific to the eastern Polynesian islands. He favored transportation via stepping stones, in both present and past times, but he noted that the progressive reduction breaks down when applied to Hawaii (1940, p. 314).

Botanists likewise have long recognized that certain of the floral elements found on existing Pacific islands represent fairly ancient stocks. Skottsberg (1928, p. 914) pointed out that because of the assumed newness of the islands they have been regarded as biological dependencies of Asia, America, or Australia. He contended that the island floras do not consist of scraps easily traced to surrounding lands. The high incidence of endemism in oceanic islands like Hawaii suggests that the flora is as old as that found on continental islands such as the Solomons and Fiji.

There is to date no paleontological evidence of great age from the large and strategically located Hawaiian islands themselves, though shallow water Cretaceous fossils have been dredged from submerged mountains less than 500 miles to the southwest. Fossils of equal age may lie buried in Hawaii. The Hawaiian mountain chain has been built up from the sea floor and it apparently has been built slowly, flow by flow, as the island of Hawaii is being built up today. This process of slow accretion is of particular interest because throughout the history of any given island most of the land is available to terrestrial life.

CENTERS OF DISPERSAL

A voluminous literature deals with the dispersal of marine invertebrate life in the Pacific. A variety of faunal provinces based, in most instances, on the distribution of some particular group of organisms has been suggested. These arrangements cannot be reviewed here, but it should be pointed out that most writers agree on two basic conceptions: (1) that connections between the Pacific and Europe via ancient Tethys existed during the Cretaceous and into Eocene time; (2) that the existing marine fauna in the islands is closely re-

lated to that of Indonesia (Edmondson, 1940; Domantay, 1953; Powell, 1958; Ekman, 1953). Many workers believe that Indonesia was the main center of dispersal; sometimes the postulated routes of migration to the islands are direct, sometimes indirect.

MEANS OF DISPERSAL

Pelagic types of marine invertebrates can drift or swim almost anywhere that currents and land barriers permit, and the same is true of organisms that encrust seaweed or other types of floating vegetation; these are the cosmopolites of the oceans. Bottom dwelling inhabitants of shallow water, on the other hand, are, in general, limited by the duration of their free-swimming larval stages. To such forms wide stretches of deep ocean may be an effective barrier. The distribution of shallow water forms among widely separated islands is, thus, not easily explained. Several agents of dispersal are considered below.

Land connections

The existing Pacific basin has been bridged by a great variety of imaginary continental lands and shallow seas, the first of which were supposed to have existed in Cambrian time (Gregory, 1930). Paleontologists were responsible for many of the ancient connections but biologists have demanded similar structures, citing the peculiarities of the existing fauna and flora. Continental drift has also been called upon to explain the existing ocean basin. In the present paper no attempt is made to go back beyond the middle Cretaceous as that is the age of the oldest fossils yet found in the island area here considered.

Data obtained by soundings, dredging, and drilling in recent years have shown that there were many more shallow banks, reefs, and high islands in the southwest part of the Pacific basin in Cretaceous and Tertiary times than there are today. Chief reason for the reduction in the numbers of these stepping stones has been subsidence that has affected large areas in the Pacific Basin proper, that is that part of the basin within the circum-Pacific andesite line (fig. 1; *see* Schmidt, 1957, p. 172-173, and Cloud, *et al.*, 1956, p. 19, fig. 1, for discussion).

In 1946 Harold Stearns called for a considerable subsidence of the Hawaiian area, citing the occurrence of coralliferous limestone in wells on Oahu more than 1,000 feet below sea level (p. 253). Also in 1946 Hess made his first report on guyots, interpreting them as wave truncated volcanoes that now lie thousands of feet below sea level. In 1954 Emery, Tracey, and Ladd (p. 152-154) summarized the evidence for subsidence in Micronesia and Polynesia. In 1956 Hamilton published evidence to show that the subsidence of some 4,000 feet took place in the area of the Mid-Pacific Mountains to the southwest of Hawaii (p. 48). Subsidence in this area led to the wide island-free gap between Hawaii and the Marshall Islands (fig. 1).

Island stepping stones

The idea that islands now sunk beneath the sea once served as stepping stones in the distribution of life to existing islands is not new. In discussing the terrestrial molluscan fauna of the Galapagos Islands, Wallace (1881, p. 270) speculated that during the long history of the existing islands some other islands may have existed between the Galapagos and the coast to serve as

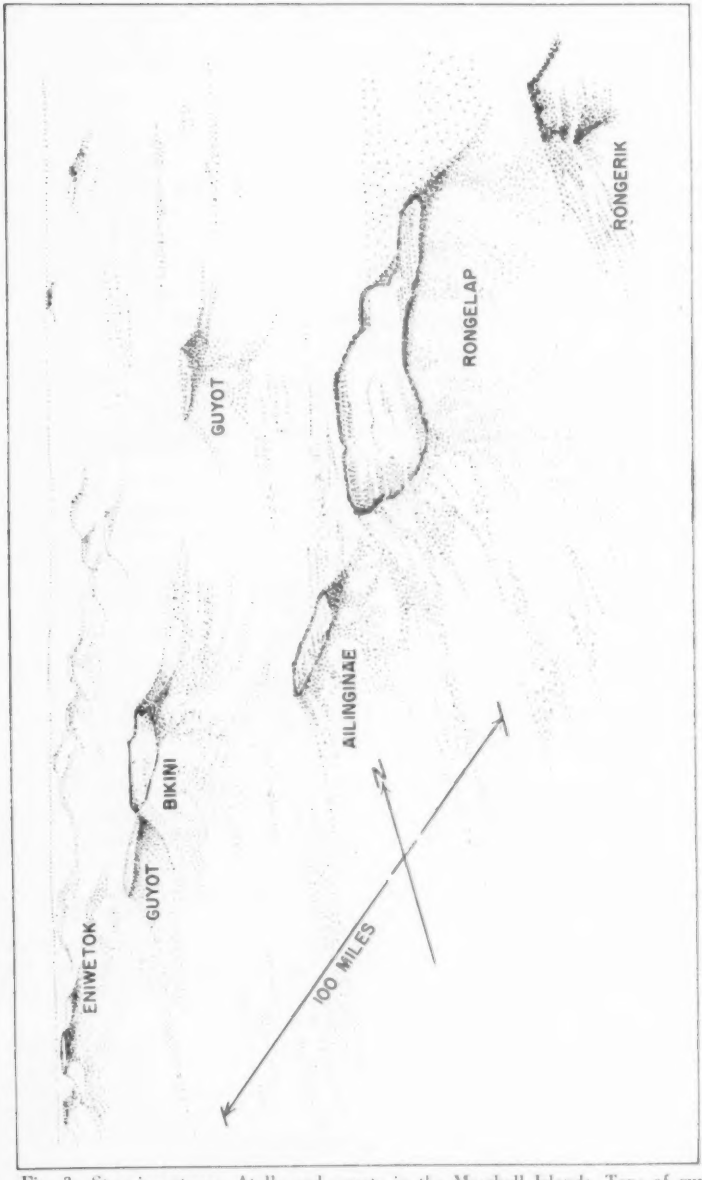

Fig. 3. Stepping stones. Atolls and guyots in the Marshall Islands. Tops of guyots, now some 4,000 feet below sea level, were truncated by marine erosion in earlier times when they projected into shallow water and aided in the distribution of marine invertebrates. Many structures of this type are present in the Mid-Pacific Mountains area (fig. 1). (After von Arx, 1954.)

"stepping stones" in the distribution of life. He noted that sunken banks, the relics of such islands, are known from many parts of the ocean and he stated that countless others no doubt remain undiscovered. Since Wallace's time, many authors, in considering the central and western Pacific, have stressed either the need for stepping stones or have given evidence for their probable existence (Vaughan, 1933, p. 933, 935; Zimmerman, 1942, p. 283; Mayr, 1953, p. 8; Hamilton, 1953, p. 206; Ladd, 1958, p. 194-196; Menard, 1959, p. 213).

Evidence from submarine mapping and dredging.—The discovery of a large number of flat-topped seamounts (guyots) scattered over millions of square miles in the western Pacific (Hess, 1946) paved the way for extensive mapping of the ocean floor and for dredging operations. One of the objectives was to test Hess' conclusion (based on configuration only) that the structures represented ancient volcanic islands truncated by wave action. Dredging on the flat summits produced rounded cobbles of basalt and a variety of shallow water fossils dating back to Tertiary and Cretaceous times (Emery, et al. 1954, p. 129; Hamilton, 1956). The pattern of distribution of the guyots is significant. A total of 50 have been confirmed by survey in the island area here considered (Menard, 1956) but only two have been discovered in the wide island-free sea that separates Hawaii and America. These two lie about 600 and 800 miles off the west coast of America (Carsola and Dietz, 1952). In this same island-free belt many volcanic seamounts are known, but none apparently projected into shallow water so as to function as stepping stones in the distribution of marine life.

Evidence from drilling.—Deep drilling on several existing atolls has demonstrated that during Tertiary and later times these structures stood well above the sea for appreciable lengths of time and could have served as stepping stones for the distribution of terrestrial mollusks, while the reefs and shallows around them continued to serve this same function for shallow water marine mollusks.

The evidence for the periodic emergence is partly petrologic, involving leaching of the limestone and the alteration of organic aragonite to calcite under atmospheric conditions (Emery, et al. 1954, p. 132); additional support is given by the occurrence of fossil land shells and concentrations of pollens and spores of land plants (Ladd, 1958). In the area to the south and west of the Hawaiian Group there are about 250 atolls plus table reefs and shallow banks that bring the total close to 300 (Bryan, 1953, Cloud, 1958).

Paleontological studies of drill cores and cuttings from the Marshall Islands show that a number of the Tertiary species (algae, Foraminifera, corals, and mollusks) that occur in outcrops above sea level outside the andesite line also lived within the Pacific Basin where they are now deeply buried below sea level.[1]

[1] At a time when additional occurrences of Tertiary rocks are being discovered in the western Pacific, an earlier occurrence should be deleted as erroneous. In 1922 Yabe and Aoki described Tertiary Foraminifera from pebbles in a reef conglomerate from Jaluit Atoll in the Marshalls, an occurrence that has been repeatedly cited in subsequent literature. Hanzawa could find no additional material on Jaluit and when he reexamined Yabe's thin sections, found two kinds of limestone. He concluded that Aoki's Recent material from Jaluit had inadvertently been mixed with Tertiary material from Saipan (Hanzawa, 1957, p. 36-37 and oral communication, 1959).

Currents

Direct transport.—In the island area outlined in figure 1 the prevailing current north of the Equator (North Equatorial Current) is to the west. South of the Equator the prevailing current (South Equatorial Current) is to the southwest and south. These two broad bands are separated by the Equatorial Counter Current flowing east. This current is stronger (more than 1 knot) but is affects only a narrow band. (Sverdrup, Johnson and Fleming, 1946, Chart 7).[2] The prevailing winds that are responsible for the currents are the Northeast Trades and the Southeast Trades, separated by a narrow, shifting belt of calms near the Equator (fig. 2).

Some of the many writers who have recognized the Indo-Pacific aspects of the island faunas have recognized also that the winds and currents set from America and cannot be called upon to support dispersal from Indonesia. W. A. Bryan pointed this out specifically for the Hawaiian marine fauna in 1921. He stated that we must look back beyond the Cretaceous for the origin of at least part of the Hawaiian fauna and flora, and he expressed the opinion that lands and currents were different in those times (p. 154-155). Edmondson likewise noted that present currents precluded direct migration from the southwest Pacific to Hawaii (1940, p. 595). Ekman noted that the wide stretch of the eastern Pacific, bare of islands, clearly forms an effective barrier to dispersal in spite of favorable currents. He also expressed the opinion that even the small and widely dispersed oceanic islands between Hawaii and Asia played a part in populating Hawaii. (Ekman, 1953, p. 21-22). Cloud, in a paper dealing with the shoal-water ecology of Saipan in the Marianas, referred to radial migration from Indonesia but pointed out that current maps for the existing ocean are not clearly reflected by biogeographic patterns. He concluded (1959, p. 396)

> Either current movements have changed recently, or are very complex or roundabout in detail, or wind-induced surface movements that run counter to or across main trends are more important than deep flow in some plankton movement. Depending on breeding seasons and wind patterns, wind-driven surface flow could be of major importance in dispersal of marine biotas.

The distribution of shallow water marine mollusks depends importantly on the duration of pelagic larval stages. Gunnar Thorson is at present engaged in a comprehensive study of this method of transport. His findings to date were summarized in a paper delivered to the International Oceanographic Congress in New York on September 9th, 1959. He reported wide variation among invertebrate groups. Among the mollusks, the pelecypods in general have a short larval period and are unsuited to long distance transport. Pelagic larvae of some prosobranch gastropods, on the other hand, have been taken alive in midocean and some of them, apparently, spend as much as 6 months drifting with the current.[3] Thorson states that 5.5 percent of mollusk larvae remain longer than 3 months in the plankton and many others remain up to 2 months.

[2] This current, however, is probably largely responsible for the Indo-Pacific aspect of the fauna of isolated Clipperton Island (Hertlein and Emerson, 1953, p. 353).

[3] All but 1 of the 27 mollusks listed by Hertlein (1937, p. 305-309) as occurring both in Polynesia and western America are prosobranch gastropods.

Under favorable temperature conditions the periods may be doubled.

In his discussion of distribution Thorson expressed the belief that currents may vary somewhat from century to century and that a maximum time for larval existence in the plankton might occur only once in 100 years.

The average speed of the equatorial currents is only about one-third of a knot (John Lyman, oral communication, 1959).[4] This amounts to 9 miles per day or a drift of some 700 miles for pelagic larvae spending 10 weeks in the plankton. Such a period would seem adequate to transport some mollusks among the stepping stones that existed in past times.

Much is still to be learned about the transportation of many pelagic larvae, but the fact remains that a variety of mollusks did spread over the scattered Pacific islands. Difficulties in moving short-lived forms are apparent but these difficulties are not nearly so great if the move is from the islands toward Indonesia rather than in the opposite direction.

Rafting.—Rafting by drifting logs may be an efficient means of dispensing marine wood-boring mollusks, and some benthonic mollusks may be distributed when the holdfasts of kelp and other sea weeds are set adrift by storms. Emery and Tschudy (1941) reviewed the literature on this subject. Mollusks and other invertebrates that cling to or are cemented to rocks may be transported long distances on floating pumice. Hedley (1899, p. 412-413) cited examples of the transportation of living reef corals by this means. Terrestrial mollusks may be rafted if, as sometimes happens, logs with attached branches, or even entire trees are uprooted and carried to sea (Ladd, 1958, p. 193-194).

Migratory birds

The migratory habits of many Pacific birds are well known (references cited in Ladd, 1958). Traffic is in both directions as far as the islands are concerned. Some birds that breed in Alaska and other boreal areas winter in the islands or in New Zealand and Australia, others that breed in New Zealand and such austral areas winter among the Pacific islands. The birds do use the islands as stepping stones and small mollusks no doubt are carried entire in particles of mud that may adhere to the feet of a bird in its flight from one island to another. Operculate snails may become locked to the feather of a bird or be eaten by a bird or fish and then regurgitated or passed with the feces (Bondesen and Kaiser, 1949, p. 268-270). The importance of such means of dispersal is difficult to evaluate but would, in any case, apply only to minute mollusks, particularly those living in shallow waters.

Typhoons

The extraordinary lifting and transporting power of typhoon winds is probably a factor of considerable importance in the distribution of small terrestrial mollusks and many types of insects (Darlington, 1957; Ladd, 1958, p.

[4] The Panamanian land bridge that now separates the Atlantic and Pacific Oceans did not come into existence until the middle Pliocene (Woodring, 1949) and it is possible that the speed of the west setting currents in the Pacific might have been higher during the Cretaceous and early Tertiary.

194). The effects of such winds in strengthening prevailing ocean currents or in developing abnormal currents may be appreciable but is difficult to evaluate.

In the northwest Pacific, tropical cyclones (including typhoons) originate north of the Equator and travel westward toward Asia where they turn northward. Similar storms in the southwest Pacific head southwestward from the equatorial regions, curving toward Australia and New Zealand (Dunn, 1951, p. 895). In 1957 there were 17 storms of typhoon intensity in the northwest Pacific (an unusually large number), one originating near Hawaii, the others in lower latitudes. If only a fourth of this number occur in the western Pacific in an average year there would have been more than half a billion typhoons in the island area since the close of the Cretaceous.

SUMMARY INTERPRETATION

The Pacific Ocean probably existed in much its present form at least as far back as the middle Cretaceous. Evidence recently obtained from soundings, dredging and drilling suggests that the southwest part of the ocean basin, in those early times, contained many more islands and shallow water banks than it does today. The shallow water marine invertebrates, including mollusks, that lived among the islands and banks could have migrated with the aid of prevailing winds and currents from the island area to the west and southwest. These early migrants may have included many elements of what is now known as the Indo-Pacific fauna.

Shallow water sediments as old as Cretaceous have recently been discovered in the Central Pacific and it is believed that existing volcanic oceanic islands, such as Hawaii, have had longer geological histories than is indicated by their meager fossil record. If, throughout their long history, they have been built up by local volcanic eruptions, there may always have been some intervening land on which ancient stocks of terrestrial faunas and floras could endure.

The Pacific island area may have been, in effect, a giant archipelago in Cretaceous and Tertiary times. Its marine fauna could migrate toward Indonesia, its terrestrial life could persist. Two heretofore conflicting lines of evidence would, thus, be reconciled—geological evidence that suggested until recently only youthful islands, and evidence from terrestrial invertebrates and plants that has long pointed to greater age.

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URANIUM AND THORIUM IN SELECTED SUITES OF IGNEOUS ROCKS*

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ABSTRACT. The uranium and thorium contents of 199 igneous rocks from a variety of petrographic provinces is here summarized. Data are given for the Mesozoic calc-alkalic batholiths of the western United States; volcanic and hypabyssal rocks of the tholeiitic magma type from Hawaii and Virginia; and effusive calc-alkalic, alkalic, and subsilicic-alkalic rocks from the western United States and Hawaii.

The batholithic rocks show an increase of both uranium and thorium from gabbro to quartz-monzonite and granite. The more extreme differentiates, chiefly muscovite-quartz monzonites, contain considerably less uranium and thorium than the quartz monzonites and granites, although the Th/U ratios are nearly the same. The volcanic and hypabyssal rocks in general show a similar increase in both thorium and uranium toward the more felsic members. The alkalic basalts of the Honolulu volcanic series show an anomalous decrease in uranium and thorium toward the right on a variation diagram.

The Th/U ratios remain fairly constant from the mafic to the felsic members of each series here studied; each series also has a more or less characteristic Th/U ratio, ranging from 2½ to 5. The scatter of thorium contents, uranium contents, and Th/U ratios is larger with increasing complexity of magmatic differentiation. The batholithic rocks have the greatest scatter, and have a significantly lower Th/U ratio in the gabbros than in the more felsic members.

Our data show no increasing loss of uranium relative to thorium from the magma during the later stages of crystallization.

INTRODUCTION

New data on the uranium and thorium contents of 199 rock samples from selected suites of intrusive and extrusive igneous rocks are here summarized. We have chosen suites of comagmatic rocks from nine regions which have been or are being studied geologically in some detail, and for all but one of which conventional rock analyses were available. Data are presented on suites of igneous rocks from the Mesozoic batholiths of the western United States: the southern California, Sierra Nevada, and Idaho batholiths. For comparison with these immense batholiths, data are also given on suites of volcanic or hypabyssal rocks from the Hawaiian Islands; Medicine Lake Highlands of California; Valles Mountains of New Mexico; Bearpaw Mountains of Montana; and the diabase-granophyre complex near Fairfax, Virginia. These suites of rocks represent petrographic provinces of widely different chemical composition and include calc-alkalic, tholeiitic, and subsilicic-alkalic magma types. All these rocks are of Mesozoic or younger age and show evidence of only minor post-magmatic alteration; the data on uranium and thorium are probably representative of the original rocks after complete solidification.

In recent years considerable data have been published on the uranium content of igneous rocks (Larsen, Jr., Phair, Gottfried, and Smith, 1956; Larsen, Jr., Gottfried, and Molloy, 1958; Adams, 1955; Whitfield and others, 1959). Data on thorium are much more scarce and little has appeared on the thorium-uranium relationships in comagmatic suites (Larsen, Jr., and Phair, 1954). Adams, Osmond, and Rogers (1959) have published an up-to-date and comprehensive review of the geochemistry of uranium and thorium which contains a comprehensive bibliography.

* Publication authorized by the Director, U. S. Geological Survey.

Adams and his colleagues have effectively applied gamma ray spectrometry to the systematic determination of thorium in ordinary rocks. In our studies, thorium has been determined by the chemical method by Levine and Grimaldi (1958) with gratifying results. The method yields results reproducible to ± 10 percent in the range above 1 ppm (part per million) thorium; the reproducibility drops off rapidly below 1 ppm to about ± 50 percent at 0.5 ppm and ± 100 percent at 0.2 ppm (the limit of determination). We have few data regarding the accuracy of the method; satisfactory agreement with gamma ray spectrometry (Adams and others, 1958) and neutron activation methods (Hamilton, 1959) has been obtained on a few samples. Uranium has been determined by the fluorimetric method of Grimaldi and others (1952) adapted to lower concentrations of uranium. The method yields results reproducible and probably accurate to ± 10 percent down to 1 ppm uranium, with increase in uncertainty to ± 100 percent at 0.1 ppm.

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We are grateful to Prof. Esper S. Larsen, Jr. who initiated these studies on uranium and continued to provide helpful advice and guidance during much of our work. We should mention particularly the enthusiastic and careful work of the U. S. Geological Survey chemists who contributed the analytical data for this report; the thorium analyses were made by Lillie B. Jenkins and Esma Campbell; most of the uranium analyses were made by Alice Caemmerer and Roosevelt Moore, and the remainder by Marion Schnepfe, Marjorie Molloy and others. Several geologists cooperated by supplying many of the rocks studied and data concerning them; they are acknowledged individually below. The variation diagrams used in this paper are patterned after those proposed by Larsen, Jr. (1938).

Part of the work reported was done on behalf of the Division of Research of the U. S. Atomic Energy Commission.

BATHOLITHIC ROCKS

Suites of igneous rocks from three large batholithic bodies, southern California, Sierra Nevada, and Idaho, have been analyzed for their thorium and uranium content. These batholiths are of calc-alkalic rocks ranging from gabbro to granite in composition. They are essentially contemporaneous in age —early Late Cretaceous according to a large number of lead-alpha age determinations (Larsen, Jr., Gottfried, Jaffe, and Waring, 1958). Larsen, Jr. and Schmidt (1958) have compared the southern California and Idaho batholiths and showed them to yield nearly identical chemical variation diagrams; the Sierra Nevada batholith is believed to be similar.

Southern California batholith

This batholith has been studied petrologically in great detail by Esper S. Larsen, Jr. (1948), and an extensive collection of the analyzed or petrographically studied specimens has been provided by him. Additional collections have been made by Gottfried. Twenty analyzed rock samples, ranging from gabbro to granite, have been further analyzed for uranium and thorium; uranium and thorium values are shown in figure 1 and the Th/U ratios in figure 10. The

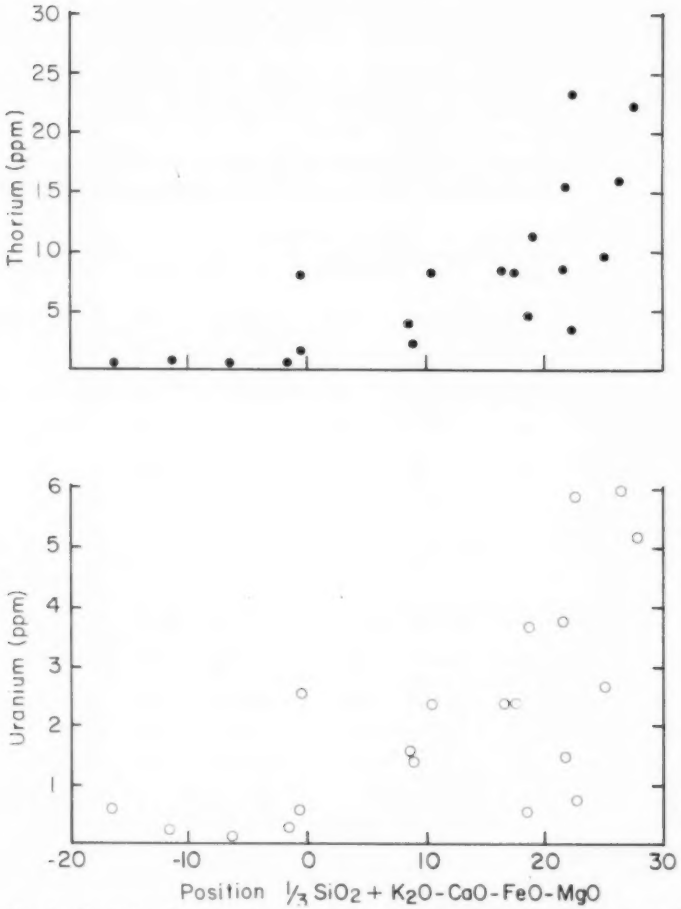

Fig. 1. Variation diagrams showing distribution of uranium and thorium in analyzed rocks of the southern California batholith.

usual relations are shown. Both uranium and thorium are quite low at the mafic end (uranium only a few tenths part per million and thorium less than one part per million in gabbro) and trend strongly upward toward the felsic end (uranium as much as 6 ppm and thorium as much as 23 ppm in the granites). The scatter of both thorium and uranium is large toward the felsic end. However, in most of those rocks in which uranium is relatively high, thorium is relatively high; when uranium is low, thorium is low. Thus the Th/U ratios show less dispersion; the mean Th/U ratio of the analyzed rocks is 3.8 and the dispersion of the individual values lie within twice or one-half of this value.

Thorium and uranium have been determined on 26 additional rock samples (studied only petrographically) and these values, together with the data on the analyzed samples mentioned above, are summarized by rock type in table 1. The wide scatter in both thorium and uranium values in each rock

TABLE 1

Thorium, uranium, and Th/U ratios in igneous rocks from the southern California, Sierra Nevada, and Idaho batholiths

Southern California

Rock types	Number of samples	Uranium (ppm) Range	Uranium (ppm) Avg.	Thorium (ppm) Range	Thorium (ppm) Avg.	Th/U Range	Th/U Avg.[1]
Gabbro	7	0.17-0.40	0.3	0.2-0.9	0.6	0.7-4.7	2.4
Tonalite	12	0.6-2.8	1.3	1.4-8.2	4.1	0.9-7.5	3.3
Granodiorite	24	1.1-4.2	2.0	3.1-15.2	7.8	2.2-6.5	4.2
Quartz monzonite and granite	2	5.2-5.3	5.2	16.0-22.2	19.1	3.0-4.3	3.7
Muscovite-quartz monzonite	1		1.0		5.5		5.5

Idaho

Rock types	Number of samples	Uranium (ppm) Range	Uranium (ppm) Avg.	Thorium (ppm) Range	Thorium (ppm) Avg.	Th/U Range	Th/U Avg.[1]
Gabbro	2	0.24-0.82	0.53	0.86-1.2	1.0	0.4-1.5	1.0
Tonalite	5	1.0-3.3	1.9	5.4-18.5	11.3	5.2-6.8	5.8
Granodiorite	12	1.1-5.1	2.4	5.4-23.2	10.3	1.6-7.1	4.8
Quartz monzonite and granite	5	2.2-6.3	4.4	14.3-22.0	17.8	2.9-8.2	4.7
Muscovite-quartz monzonite	5	0.5-1.9	1.5	4.9-7.9	6.5	4.7-5.8	5.3

Sierra Nevada

Rock types	Number of samples	Uranium (ppm) Range	Uranium (ppm) Avg.	Thorium (ppm) Range	Thorium (ppm) Avg.	Th/U Range	Th/U Avg.[1]
Granodiorite	2	5.2-5.2	5.2	14.3-24.0	19.2	2.8-4.7	3.8
Quartz monzonite and granite	5	2.3-7.2	4.3	7.6-36.0	18.7	3.3-5.0	4.2
Albite-quartz monzonite	1		2.0		9.4		4.8

[1]These data are the averages of individual Th/U ratios, not the ratio of the average thorium and uranium contents.

type is apparent, but the average thorium and uranium values (the arithmetic means) increase regularly from gabbro toward granite. The Th/U ratios show a somewhat greater dispersion for each rock type than for the analyzed rocks alone, perhaps owing to the inclusion of more erratic and uncommon rock types. Considering the batholith as a whole, the gabbros have a slightly lower Th/U ratio than the later differentiates; within the uncertainties of the averages, the rocks between tonalite and granite have much the same average Th/U ratio—between 3½ and 4.

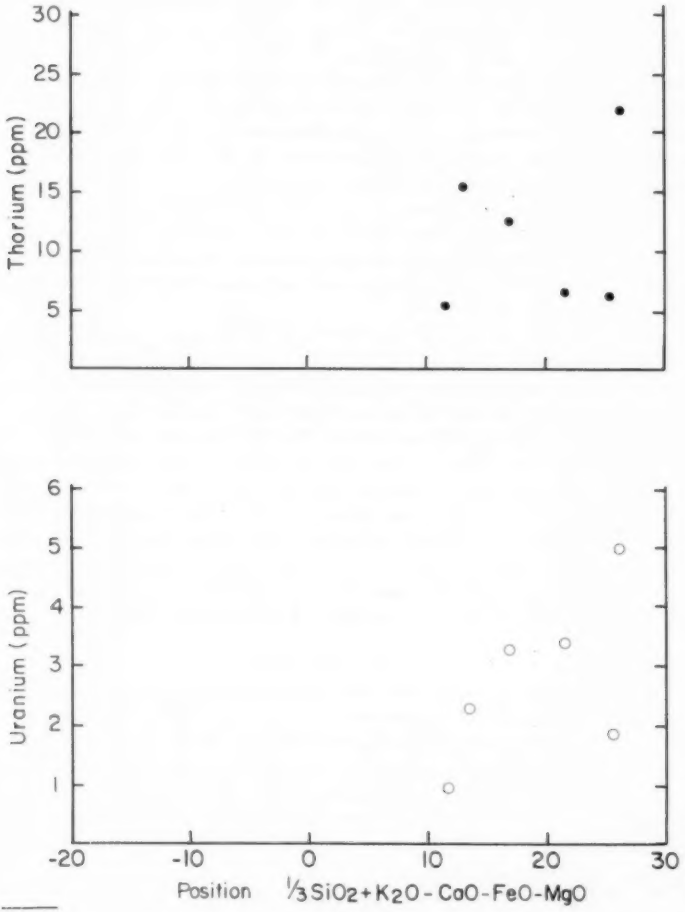

Fig. 2. Variation diagrams showing distribution of uranium and thorium in analyzed rocks of the Idaho batholith.

Idaho batholith

The Idaho batholith has been studied petrologically in broad reconnaissance by Larsen, Jr., and Schmidt (1958), and their collections of samples have been made available for this study. Only six analyzed rock samples, all of relatively felsic types, were available for uranium and thorium determinations; these data are shown on variation diagrams in figure 2. The wide scatter of both thorium and uranium values is quite similar to that shown for the equivalent analyzed rocks from the southern California batholith (fig. 1) and the Sierra Nevada batholith (fig. 3, see below). Table 1 presents the range and average thorium and uranium values for 29 rocks tabulated by rock type. The range of uranium values and the average uranium content for each rock type are very nearly the same (considering the relative uncertainties introduced by small numbers of samples and the analytical uncertainties below 1 ppm uranium) as those for the southern California batholith. The thorium values for gabbro and for quartz monzonite and granite are nearly the same as for comparable rocks from southern California, but the thorium contents of the tonalites and granodiorites are significantly greater. The Th/U ratios of the gabbros are low and comparable to those of the southern California rocks (fig. 10), but the Th/U ratios of the tonalites and more silicic rocks are significantly higher —the mean Th/U ratio ranging roughly between 4½ and 5½. Nearly all the samples have a Th/U ratio between twice and one-half the tabulated averages.

Sierra Nevada batholith

Eight analyzed samples from the eastern part of the Sierra Nevada batholith, in the Bishop area, were supplied by Paul Bateman, of the U. S. Geological Survey. These samples are of quite limited petrographic type, ranging from granodiorite to albite-quartz monzonite. The thorium and uranium contents of each are plotted on variation diagrams in figure 3. They show no relation of thorium or uranium to rock composition, but rather show a wide dispersion of both thorium and uranium, even greater than that of similar rocks from the southern California and Idaho batholiths. The ranges and average values of uranium and thorium, and of the Th/U ratios are tabulated for comparison in table 1. The range of Th/U ratios is quite small and the average values are near those of similar rocks in the southern California batholith.

Late differentiates

A few samples of rather extreme petrographic composition, and believed to be quite late in the intrusive history of the batholiths, have been analyzed for uranium and thorium. The analytical data are shown in table 1 for a single sample of muscovite-quartz-monzonite from the southern California batholith; five similar samples from the Idaho batholith; and a single sample from the Sierra Nevada batholith. They are markedly lower in both uranium and thorium than the more typical quartz-monzonites and granites. Paradoxically, the relatively rare accessory minerals monazite, xenotime, and highly radioactive zircon are commonly found in these muscovite-bearing rocks. The five muscovite-quartz monzonites from the Idaho batholith are particularly convincing in their low contents of both elements. The Th/U ratios, however, are not sig-

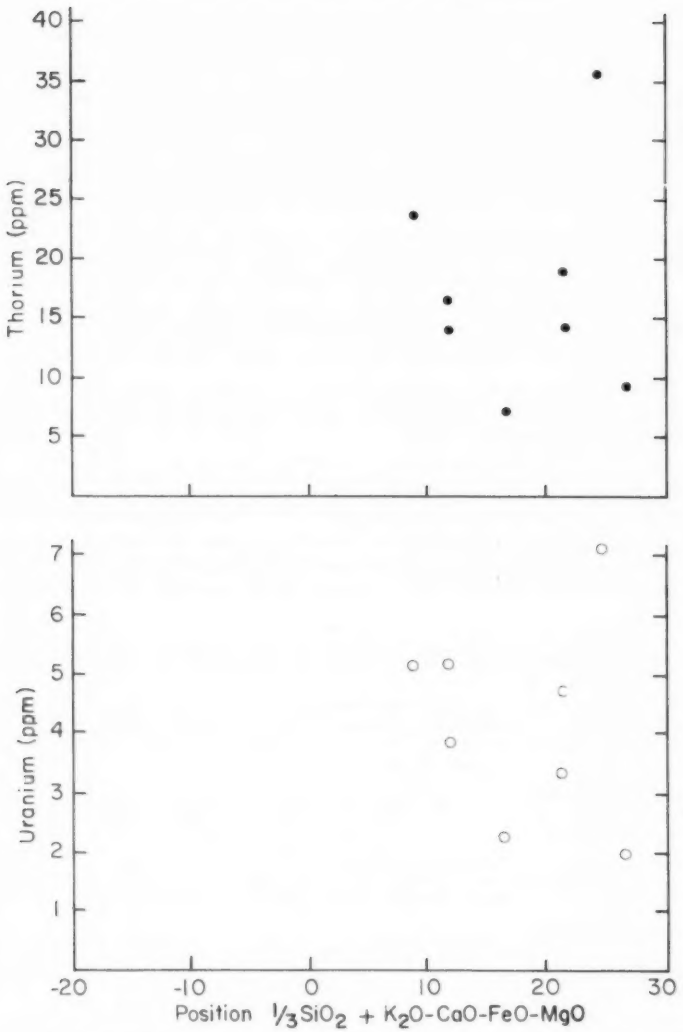

Fig. 3. Variation diagrams showing distribution of uranium and thorium in analyzed rocks of the Sierra Nevada batholith, California.

nificantly different from the more common rocks of these batholiths. Clearly neither uranium nor thorium has been selectively removed, but both uranium and thorium have been depleted in or lost from these "residual" magmas in constant proportions. We have no ready explanation for this.

VOLCANIC AND HYPABYSSAL ROCKS

Hawaiian Islands

Uranium and thorium were determined on 35 samples of volcanic rocks believed to be fairly representative of the major rock types of the Hawaiian Islands. Olivine basalts are the predominant rock type and according to Macdonald (1949) they represent the parent magma of the Hawaiian petrographic province. Powers (1955) points out that the olivine basalts of the older primitive shields are rarely silica deficient. He suggests that the name "olivine basalt" not be applied to the magma as it suggests a magma undersaturated with silica.

The undersaturated rocks that are present occur in considerably smaller amounts than the ordinary Hawaiian basalts and were formed during the declining stages of activity at some of the volcanoes.

Tilley (1950) showed that the basalts of the primitive shields from the Hawaiian Islands are very similar chemically to great basaltic provinces on the continents, such as the Triassic basalts from New Jersey, the Karoo dolerites, and the Deccan basalts of India.

All the samples which we have analyzed for uranium and thorium were made available by Howard Powers of the U. S. Geological Survey. These include samples studied by him and by other workers. A chemical rock analysis is available for each of the samples. The uranium and thorium values are summarized in table 2. Basalts from the Koolau volcanic series of Oahu studied by Wentworth and Winchell (1947) contain larger amounts of silica than most

TABLE 2

Uranium and thorium contents of volcanic rocks from the Hawaiian Islands

Location	No. of samples	Uranium Range	Uranium Avg.	Thorium Range	Thorium Avg.	Th/U Range	Th/U Avg.[1]
Basalts of primitive stage of eruption							
Koolau volcanic series, Oahu	7	0.27-0.54	0.32	0.53-1.2	0.79	1.8-4.5	2.6
Mauna Loa, Hawaii	7	.21-0.28	.25	.63-0.98	.80	2.3-4.1	3.2
Kilauea, Hawaii	9	.28-0.41	.35	.84-1.4	1.1	2.6-3.5	3.1
Volcanics of declining stage of eruption							
Andesite, Mauna Kea, Hawaii	1		1.3		3.6		2.8
Oligoclase andesite, Kohala	1		1.2		5.3		4.3
Olivine basalt, Kohala	1		.49		1.3		2.7
Olivine basalt, Hualalai, Hawaii	1		.41		1.4		3.4
Picrite basalt, Haleakala, Maui	1		.45		2.0		4.3
Alkalic basalts, Honolulu volcanic series, Oahu	7	0.58-3.2	1.8	2.5-10.4	5.4	2.6-4.7	3.1

[1] These data are the averages of individual Th/U ratios, not the ratio of the average Th and U contents.

Hawaiian basalts. They average about 0.3 ppm uranium, 0.8 ppm thorium, and have an average Th/U ratio of 2.6.

The samples from Mauna Loa and Kilauea are mostly olivine basalts. The uranium and thorium contents of the samples from Mauna Loa are exceedingly uniform. They average 0.25 ppm uranium and 0.8 ppm thorium. The Th/U ratio is 3.2. The samples from Kilauea also show marked uniformity in their uranium and thorium contents and have nearly the same average uranium and thorium contents as the samples from Mauna Loa. The average Th/U ratio for the Kilauea samples is 3.1.

The samples of the lavas erupted during the late, declining stage of volcanic activity form a thin veneer over the earlier extrusives. The volume of the late-stage volcanics is insignificant as compared to the volume of the earlier formed basalts. The basalts of the Honolulu volcanic series overlie the basalts

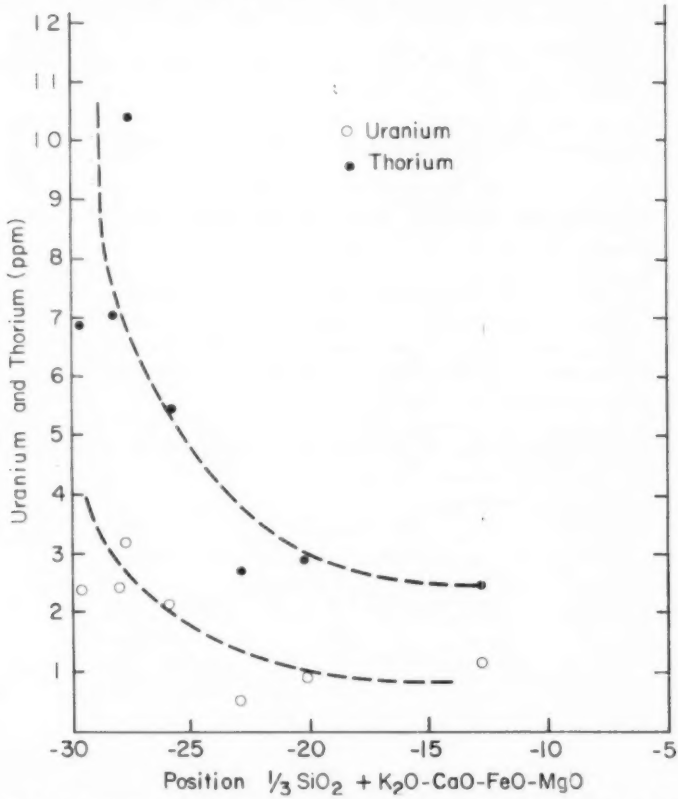

Fig. 4. Variation diagram showing distribution of uranium and thorium in analyzed rocks of the Honolulu volcanic series, Oahu, Hawaii.

of the Koolau volcanic series on the island of Oahu and are separated from them by a long erosion interval. They are low in silica (36-45 percent) with a relatively high alkali content. Petrographic studies show them to be nepheline basalt, nepheline basanite, and nepheline-melilite basalt (Winchell, 1947). The uranium and thorium values of seven samples from this series are plotted on a variation diagram on figure 4. Unlike any of the other variation curves showing the relation of uranium and thorium to chemical composition, the curves for the rocks of the Honolulu series show a decrease of uranium and thorium towards the right side of the diagram. Their average uranium (1.8 ppm) and thorium content (5.4 ppm) is higher than for any other series of volcanics from the Hawaiian Islands. However, the average Th/U ratio 3.1 is very similar to the average for the Mauna Loa and Kilauea samples (fig. 10).

Uranium and thorium determinations have been made on five samples from "late" eruptions. The two andesites contain more uranium and thorium than any of the olivine basalts. The three "late" basalts from Kohala and Hualalai on Hawaii, and Haleakala on Maui, contain a little more than 0.4 ppm uranium and 1.6 ppm thorium, which are a little higher than the average contents of the ordinary basalts. The average Th/U ratios of the two andesites is 3.5 and of the olivine basalts 3.6. More samples from these late eruptions are needed, however, before any firm conclusions can be drawn.

Fairfax County, Virginia

Twenty samples of igneous rocks from the diabase-granophyre sequence were analyzed for uranium and thorium contents. Because of lack of chemical data, the rocks are grouped in three categories based on petrographic data (E. C. T. Chao, U. S. Geological Survey, oral communication, 1958); (1) typical diabase, which consists essentially of monoclinic pyroxene and labradorite; (2) an intermediate phase containing amphibole, less calcic plagioclase, and small amounts of quartz; and (3) granophyric rocks consisting of albitic plagioclase, amphibole, and quartz.

The uranium and thorium determinations are summarized in table 3. Uranium and thorium show nearly a fivefold enrichment in the granophyric rocks over the typical diabase. The average Th/U ratios are approximately the same for each of the rock types.

TABLE 3

Uranium and thorium content of diabase-granophyre rocks[1] from Fairfax, Virginia

Rock type	No. of Samples	Uranium Range	Uranium Average	Thorium Range	Thorium Average	Th/U[2]
Typical diabase	11	0.2-0.7	0.5	0.8-3.1	2.1	4.3
Intermediate phase	5	0.8-2.0	1.4	2.6-8.5	5.4	3.9
Granophyric	4	1.9-3.3	2.5	8.2-13.1	9.6	4.0

[1] Samples supplied by E. C. T. Chao, U. S. Geological Survey.

[2] These data are the averages of individual Th/U ratios; not the ratio of the average uranium and average thorium contents.

Medicine Lake Highland, California

The Medicine Lake Highland is located approximately in the center of the Modoc Lava Bed quadrangle. The lavas which make up the Highland represent a well-differentiated series from basalt to rhyolite. The eight samples analyzed for uranium and thorium for this report are parts of the same material studied

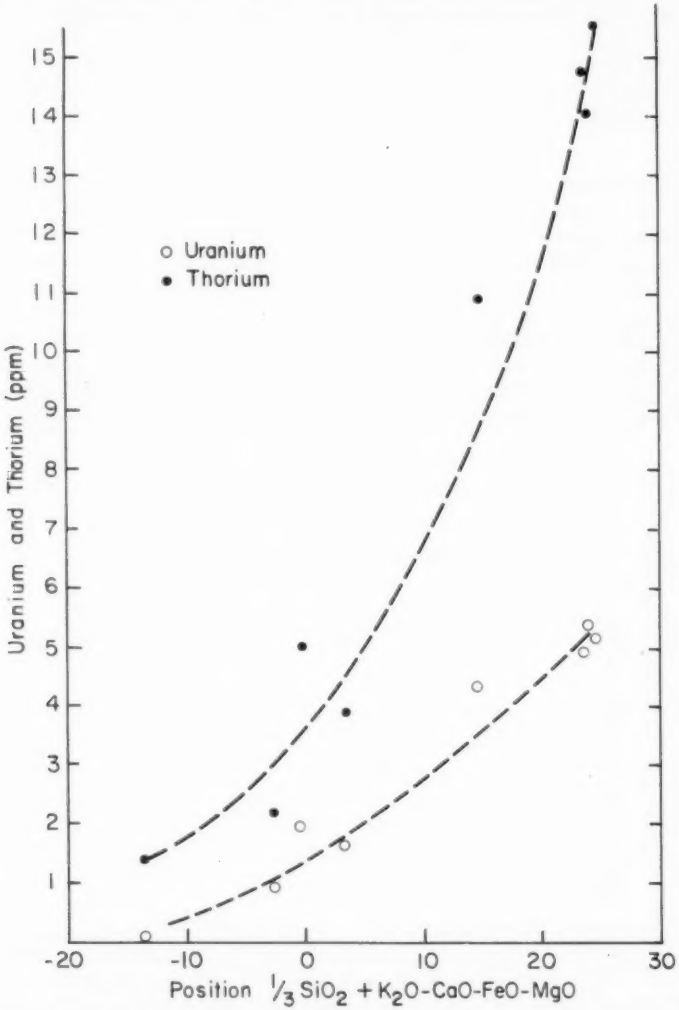

Fig. 5. Variation diagram showing the distribution of uranium and thorium in analyzed rocks of the Medicine Lake Highland, Calif.

by Anderson (1941), and were supplied by William Quaide of the Department of Geology, University of California (Berkeley). Conventional rock analyses are available for each specimen; the relation between chemical composition and uranium and thorium contents is shown in figure 5, and the Th/U ratios in figure 10. Both uranium and thorium are present in greater amounts in the more siliceous rocks than in the mafic rocks. One sample of basalt is extremely low in uranium, containing about 0.1 ppm (based on replicate analyses), and having a Th/U ratio of 14. The Th/U ratios in the two other basalts are 2.2 and 2.5. The rhyolites have Th/U ratios of 2.7 to 3.0. Except for one basalt with the Th/U ratio of 14, there seems to be a small increase in the Th/U ratios from basalts to rhyolites.

Lassen Volcanic National Park, California

The volcanic rocks from Lassen Volcanic National Park range from basalt to dacite. Compared with the volcanic rocks from the Medicine Lake Highland, which is regarded as a calc-alkalic series, the Lassen volcanic rocks are regarded as a calcic suite. Adams (1955) has reported uranium determinations and alpha activity on 39 samples from this area. He found that uranium increases with K_2O and that the relative alpha activity to uranium ratio is constant, indicating a constant Th/U ratio. Uranium and thorium determinations on 9 samples of volcanic rocks studies by Williams (1932) are plotted on a variation diagram on figure 6. These samples were supplied by William Quaide of the Department of Geology, University of California (Berkeley). There is a

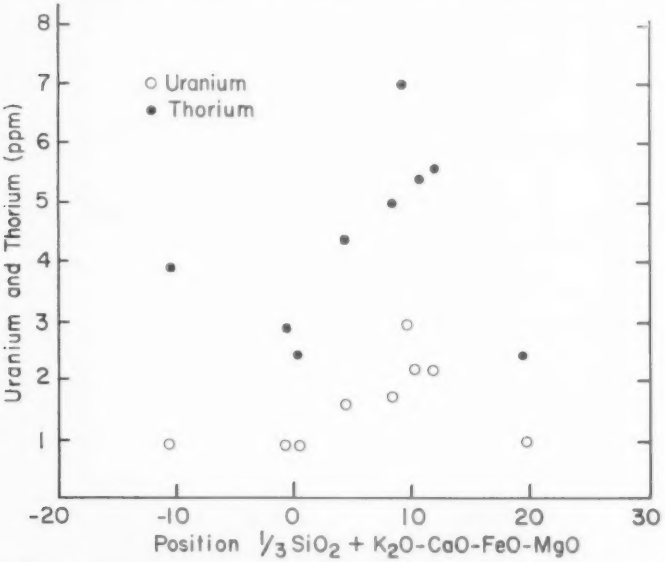

Fig. 6. Variation diagram showing the distribution of uranium and thorium in analyzed rocks of the Lassen Volcanic National Park, Calif.

clear trend toward increasing uranium and thorium in the more siliceous rocks except for one sample of dacite which is low in both uranium and thorium. This sample of dacite, the most siliceous rock of the suite analyzed for uranium and thorium, contains only 1.0 ppm uranium and 2.4 ppm thorium.

Too few samples have been analyzed by us to show any definite correlation of Th/U ratios with rock type. However, the available data indicate a small decrease in the Th/U ratios from basalt to dacite (fig. 10). The two basalts

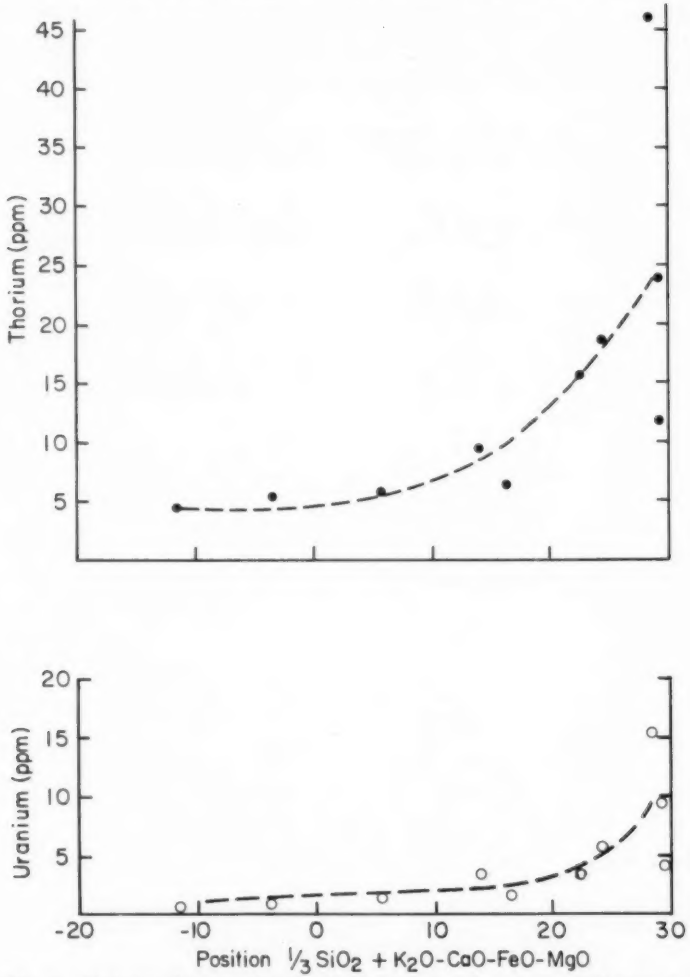

Fig. 7. Variation diagrams showing the distribution of uranium and thorium in analyzed rocks of the Valles Mountains, N. Mex.

have Th/U ratios of 4.0 and 3.0; the average Th/U ratio of five andesites is 2.7; and in the two dacites the Th/U ratio is 2.7 and 2.4.

Valles Mountains, New Mexico

A suite of 10 samples from the Valles Mountains consisting of basalt, andesite, dacite, rhyodacite, and rhyolite were analyzed for uranium and thorium. The samples and chemical analyses for each sample were provided by Robert L. Smith and Roy Bailey (U. S. Geological Survey). This suite of volcanic rocks is somewhat more alkalic than the Lassen and Medicine Lake Highland volcanic rocks. The data (fig. 7) show a fairly systematic increase of

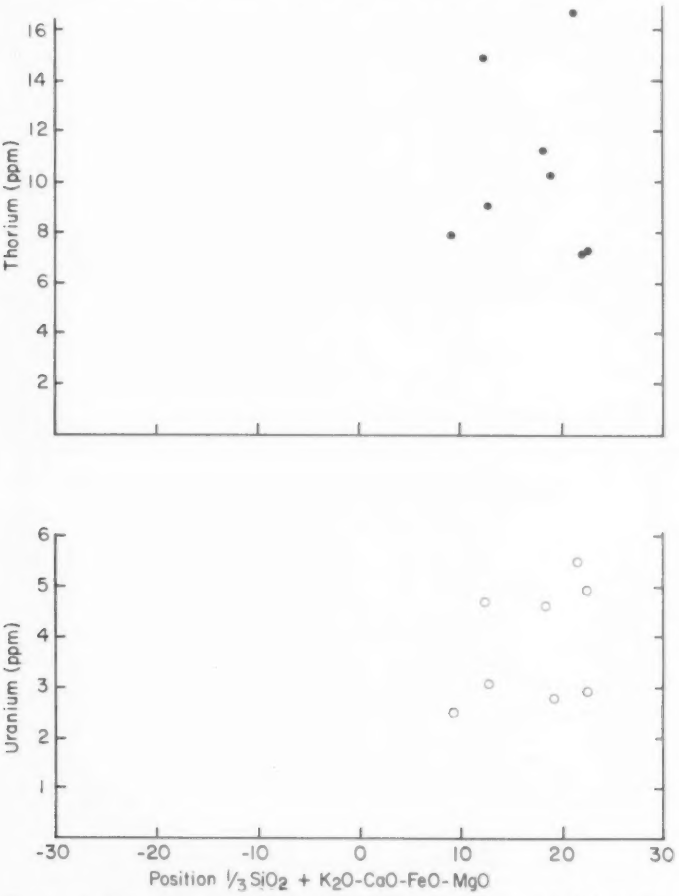

Fig. 8. Variation diagrams showing the distribution of uranium and thorium in analyzed silicic rocks of the Bearpaw Mountains, Mont.

uranium and thorium from the mafic to the siliceous rocks. The basalts contain about 1.0 ppm uranium and about 5.0 ppm thorium. The rhyolites range from 4 to about 15 ppm uranium and from 12 to 46 ppm thorium. The increase in uranium and thorium from basalts to rhyolites is greater than for any of the other volcanic provinces discussed in this report.

The Th/U ratios (fig. 10) show no systematic change with composition. The average Th/U ratios are as follows: basalt, 5.5; andesite, 3.3; dacite, 2.6; rhyodacite, 4.1; and rhyolite, 3.0.

Bearpaw Mountains, Montana

Thirty-four samples of extrusive and intrusive igneous rocks from the Bearpaw Mountains were analyzed for uranium and thorium. Eight of the samples are different types of latitic rocks from a silicic series, and 26 samples are rocks of a subsilicic-alkalic series which are predominantly mafic phono-

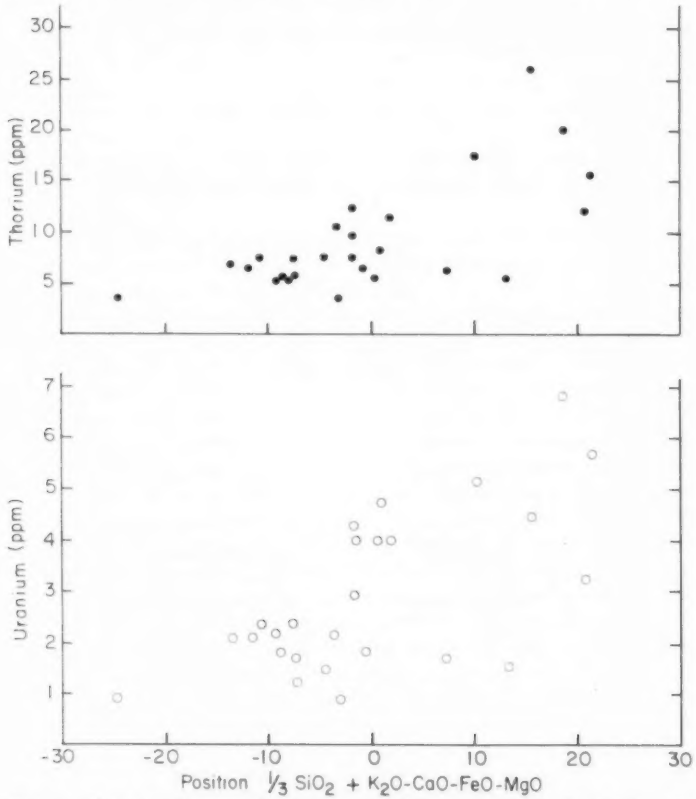

Fig. 9. Variation diagrams showing the distribution of uranium and thorium in analyzed subsilicic-alkalic rocks of the Bearpaw Mountains, Mont.

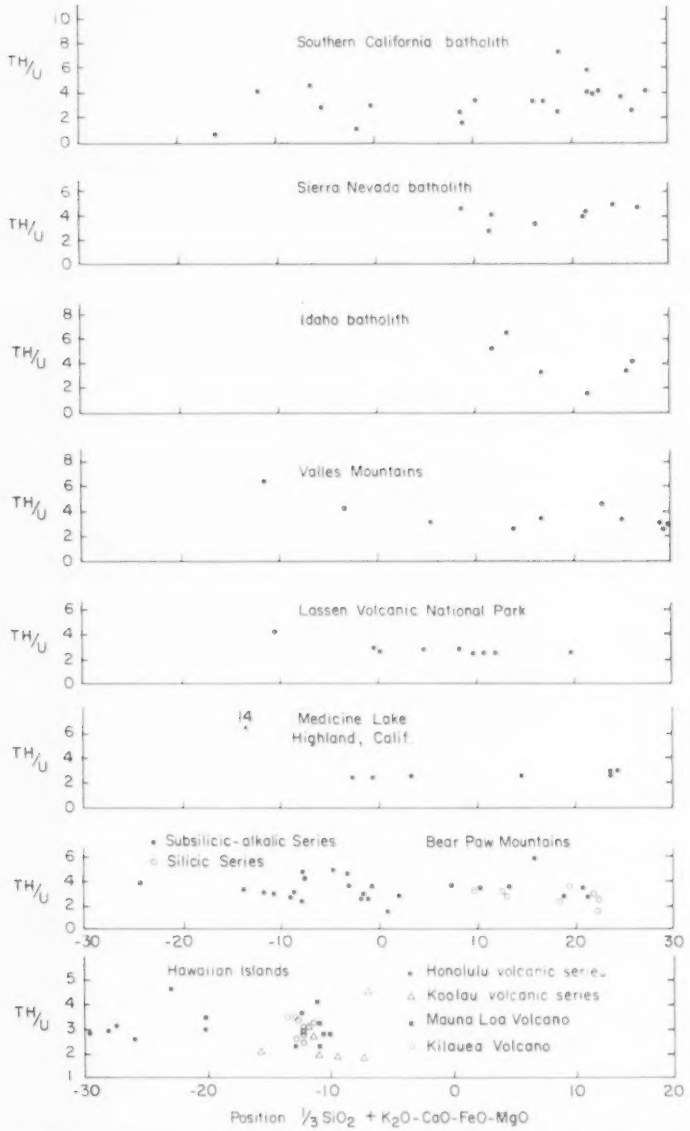

Fig. 10. Variation diagrams showing the distribution of thorium-uranium ratios in analyzed rocks from eight of the igneous provinces considered in this report. Note that the horizontal and vertical scales of the bottom diagram (Hawaiian Islands) are different from the others.

lites and shonkinites. The samples and chemical analyses were made available to us by W. T. Pecora and Robert G. Schmidt, U. S. Geological Survey. The latitic rocks average 3.9 and 10.6 ppm uranium and thorium respectively, and have an average Th/U ratio of 2.8. The uranium and thorium data for these samples are plotted on figure 8 and the Th/U ratios on figure 10.

Figure 9 shows the distribution of uranium and thorium in 26 rocks of the subsilicic-alkalic series. The most mafic rock is a biotite pyroxenite which contains a little less than 1 ppm uranium and 3.9 ppm thorium. The shonkinites and phonolites average 2.4 and 7.2 ppm of uranium and thorium respectively. The syenites, which lie on the right side of the variation diagram, average 4.1 and 14.3 ppm uranium and thorium. The Th/U ratio of each of the rocks is plotted in figure 10. The average Th/U ratio for the biotite pyroxenite is 4.1· the shonkinites and phonolites, 3.3; and the syenites 3.6.

DISCUSSION AND CONCLUSIONS

1. In the calc-alkalic and subsilicic-alkalic series of rocks we have studied, there is the usual increase in both uranium and thorium from the mafic to the felsic types. The scatter of both uranium and thorium values from "average" curves for each series is greater the more complex the history of differentiation of the magma producing the rocks. The batholithic rocks show an increasing dispersion of these elements toward the felsic end, whereas several of the volcanic series yield "average" curves from which analyzed values deviate very little.

The Honolulu volcanic series, a series of undersaturated basalts very low in SiO_2 and relatively high in alkalies, shows an unusual and strong reversal of the trend of both uranium and thorium in a variation diagram. Uranium and thorium decrease markedly and uniformly toward the right on a variation diagram (fig. 4), although they do not vary systematically with K_2O.

2. Several of the igneous provinces or subprovinces studied have more or less characteristic variation curves for both uranium and thorium. Provinces having relatively greater alkali contents but normal silica contents (such as the Valles Mountains) contain more of both thorium and uranium, on the average.

3. The Th/U ratios in the igneous rock series considered here are rather uniform for each series, although each series has a more or less characteristic ratio ranging approximately from $2\frac{1}{2}$ to 5. In the simpler volcanic series, the ratios are fairly constant from basalts to rhyolites. In the volcanic series believed to have a more complicated pattern of differentiation, the Th/U ratios show somewhat more dispersion, but the average Th/U ratios are still relatively constant, within the limits of sample variations and analytical errors, regardless of stage of differentiation. In the batholithic rocks, the gabbros are significantly low in thorium relative to uranium, yielding low Th/U ratios. From the tonalites to the granites, the mean Th/U ratio of each rock type is essentially constant. This is true even of the late, more extreme differentiates, which are relatively low in both thorium and uranium, but show quite "normal" Th/U ratios.

Our data are in conflict with those of Whitfield and others (1959). They report an aparent increase in thorium relative to uranium during petrogenetic

evolution (Whitfield and others, 1959, p. 264). Except in the gabbros of the batholithic rocks, we find no systematic variation of thorium relative to uranium; rather, the two elements, on the average, remain in remarkably constant proportions for each province. However, there are clear provincial differences; for example, the tonalites to granites of the southern California batholith have a mean Th/U ratio near 4, whereas similar rocks of the Idaho batholith yield a ratio near 5.

In late-stage differentiates which are associated in space and in time with vein deposits of pitchblende, Phair (1952) found that both thorium and uranium are greatly enriched, but that thorium is more enriched than uranium, the Th/U ratios averaging about 7½. He believes that, for these rocks, the relative enrichment of thorium was the result of a loss of some uranium from the late differentiates because of oxidation of uranium and its consequent increased solubility. From their studies of scattered samples (and not of petrogenetic suites), Whitfield and others (1959) have drawn the more far-reaching conclusion that increasing Th/U ratios during petrogenetic evolution of granitic rocks are the general rule, and that oxidation of uranium and its increased solubility is the usual mechanism for this relative enrichment of thorium. Our data, however, show no evidence of such systematic loss of uranium from the magma (or increase of thorium in the rocks) during the later stages of differentiation of the igneous provinces reported on here. We are unable to accept the suggested mechanism to explain the relatively greater abundance of uranium in the gabbros of the batholiths although we have no other explanation to offer.

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GEOLOGIC PROBLEMS IN THE SWEETWATER BARITE DISTRICT, TENNESSEE*

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ABSTRACT. Residual barite deposits, derived from vein and breccia fillings in carbonate rocks of Ordovician age have been mined extensively in three parallel strike belts near Sweetwater, Tennessee. Geologic problems germane to the further exploitation of this mineral resource include the origin and age of the primary deposits and the breccias which contain them; relation of the barite deposits to zinc deposits of East Tennessee; weathering of the carbonate rocks, including accumulation of the residual mantle; solution, transportation, and deposition of barite, fluorite, and silica; and the extent of ancient terrace deposits. The district presents a promising field for coordinated research in economic geology, geochemistry, geophysics and geomorphology.

INTRODUCTION

Barite occurs at many localities in the Valley and Ridge physiographic province from Pennsylvania to Alabama; most of the commercial production, however, has come from two major districts—Cartersville, Georgia and Sweetwater, Tennessee, with minor production from many scattered districts and localities.

Barite in Tennessee was first reported in 1818 (Kain, 1818, p. 63) from Sevier County, but the earliest commercial production in the state appears to have been about 1870 from the Sweetwater district (Gordon, 1918, p. 72). However, galena, with barite and fluorite which were discarded as gangue minerals, was mined at the Carter mine, three miles east of Sweetwater, as early as 1826 (Safford, 1869, p. 485).

The Sweetwater district (fig. 1) is between the Tennessee and Hiwassee

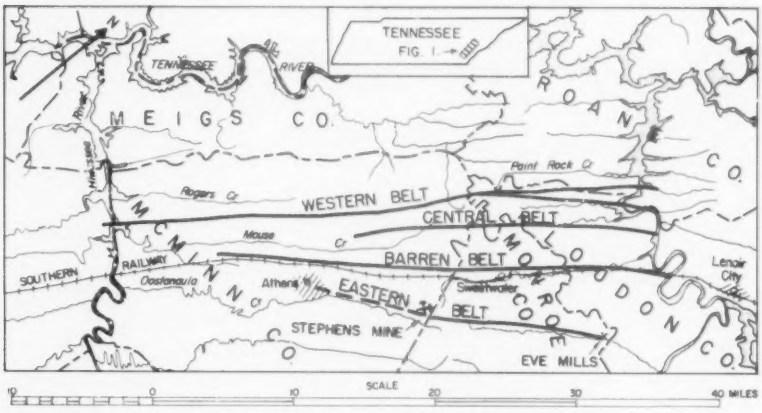

Fig. 1. Map of the Sweetwater barite district.

Rivers, in Monroe, McMinn, Loudon and Roane Counties, Tennessee, about half way between the cities of Knoxville and Chattanooga, and about 50 miles southwest of the Mascot-Jefferson City zinc district. The district is traversed

* Publication authorized by the Director, U. S. Geological Survey.

by U. S. Highway 11 and a main line of the Southern Railway System. Although early records are incomplete, it is reasonably certain that about one and a half million tons of barite were shipped from the Sweetwater district between 1870 and 1959 and that approximately half of this total came from the Stephens mine in the southeastern part of the district.

The earliest report of barite production in the Sweetwater district was by Killebrew and Safford (1874, p. 270). Curiously, no mention is made of the barite deposits in the Kingston and Loudon folios (Hayes, 1894; Keith, 1896), although most of the district is in these two quadrangles. Probably this omission resulted from the fact that the district was almost completely unproductive between 1875 and 1899 (Gordon, 1918, p. 72). Other early descriptions of the district were by Judd (1907), Fay (1909), Henegar (1912), Grasty (1913), and Watson and Grasty (1915), but most of these were more concerned with mining and milling methods than with geology or exploration.

In 1918 Gordon published a paper which summarized the geology of the district and included descriptions of 22 mines and prospects. This report remains the most complete geological report on the barite deposits of the Sweetwater district. Later papers, dealing chiefly with the origin and structural control of the primary deposits, were those of Secrist (1924), Laurence (1939), Dunlap (1945, 1955) and Kesler (1950). Penhallegon (1938), Rankin and others (1938) and Gildersleeve (1946) described mining and concentration, with brief descriptions of the deposits. Part of the district was mapped geologically by Rodgers (1952, 1952a), who also provided a brief description of the barite deposits in one of the quadrangles (1952). The most recent publication is a brief summary by Brobst (1958). Exploration for zinc sulfide deposits was carried on by two mining companies in the early 1950's, and at least two companies have recently done some systematic exploration for residual barite deposits, but the results of these projects have not been made public.

GEOLOGY OF THE DISTRICT

The Sweetwater district is underlain by more than 7,000 feet of sedimentary rocks of Lower Cambrian to Middle Ordovician age, which have been folded and faulted so that they are in several separate outcrop belts (Rodgers, 1952). The barite deposits are in the Knox group, which has been subdivided as shown in table 1 (Rodgers, 1952; Dunlap, 1947):

TABLE 1

Formations of the Knox group

Formations	Thickness (feet)
Lower Ordovician	
Mascot dolomite	500 - 600
Kingsport formation	280 - 325
Upper member (dolomite)	150
Lower member (limestone)	130 - 175
Longview dolomite	250 - 300
Chepultepec dolomite	750
Upper Cambrian	
Copper Ridge dolomite	1000

With few exceptions, these formations strike northeast and dip less than 30 degrees to the southeast; they form several blocks separated by low-angle thrust faults which dip to the southeast. These faults, whose traces traverse the district from northeast to southwest, are major structural features which extend far beyond the district (Rodgers, 1953). From southeast to northwest, the major thrust faults are the Chestuee, Knoxville, Saltville, Beaver Valley, and Copper Creek. Their relationship to the major subdivisions of the Sweetwater barite district is shown diagrammatically in figure 2.

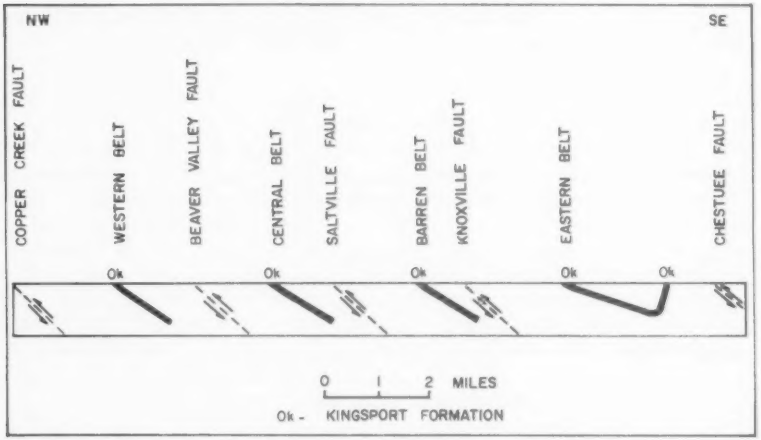

Fig. 2. Diagrammatic cross section of the Sweetwater barite district.

The barite deposits occur in three linear belts, mainly along the outcrop of the Kingsport formation. Gordon (1918, p. 55) called these areas the Howard, Garrison, and Culveyhouse veins but now they are commonly known as the eastern, central, and western belts. The northern part of the western belt forks to form two belts, due to repetition of the Kingsport formation by a minor fault asociated with the Beaver Valley thrust fault. Between the eastern and central belts is another outcrop belt of the Kingsport formation, which is not known to contain any barite deposits and is therefore called the barren belt. There are also four belts of the Kingsport formation to the southeast and four to the northwest of the district; only a few minor barite prospects are known in those to the southeast and none at all in those to the northwest.

The rocks of the district have been subjected to weathering and erosion since the close of the Paleozoic era, and this has produced the typical features of the Knoxville-Chattanooga segment of the southern section of the Valley and Ridge physiographic province (Fenneman, 1938, p. 271-272). These are two prominent levels, the upper one (Harrisburg peneplain) at 1,200 to 1,400 feet above sea level on ridges held up by cherty dolomites and sandy shales, and the lower one (Coosa peneplain), which lies at an elevation of 800 to 900 feet in the alluviated valleys underlain by non-cherty limestones and calcareous

shales. Throughout much of the area of carbonate rocks, bedrock is covered by residuum, which ranges from a few feet to more than 150 feet thick, and natural exposures, especially of the Knox group, are relatively scarce. Evidence of old, high-level alluvial terraces is seen in the presence of scattered cobbles of quartzite and vein quartz derived from Cambrian and Precambrian rocks of the Blue Ridge province.

THE BARITE DEPOSITS

All commercial production from the Sweetwater district has been from residual deposits, but primary deposits are exposed in the bedrock in the deeper parts of several of the mines. In fact, no mine which has been worked down to bedrock has failed to expose primary deposits. These occur as veins in intensely brecciated beds of dolomitized limestone, similar to the well-known "recrystalline" of the Mascot-Jefferson City zinc district (Oder and Hook, 1950, p. 75-81). No large, persistent veins are known. Generally only a few inches in width, the veins commonly expand at intersections into large, irregularly shaped masses, and elsewhere thin into mere stringers. The contact between vein and wall rock is commonly sharp. The principal minerals of the veins are barite, fluorite, and pyrite, with minor galena, sphalerite, and calcite. Fluorite and galena are apparently more common in the eastern belt and the northern part of the western belt, but this may be simply the result of better exposures in those places. A large unmined zinc deposit occurs at Eve Mills, at the northeastern end of the eastern belt (Dunlap, 1947).

Almost all the exposed deposits are in the limestone member of the Kingsport formation, and most of the residual deposits are in residuum derived from that member. Table 2 shows the stratigraphic distribution of 112 mines and prospects, partly from data by Rodgers (1952) and an unpublished manuscript by J. C. Dunlap. It should be noted that these deposits are in the residuum of the formation indicated.

TABLE 2

Stratigraphic distribution of barite deposits

Formation	Eastern Belt	Central Belt	Western Belt
Mascot	6	4	7
Kingsport (upper member)	8	6	6
Kingsport (lower member)	44	4	12
Longview	3	1	3
Chepultepec	4	0	0
Copper Ridge	0	0	0

The residual deposits consist of nodules and larger masses of barite, chert and limonite in reddish or yellow clay. Individual masses or "dornicks" of barite may measure as much as two feet in the longest dimension, although they are commonly less than six inches long. Barite-bearing clay may crop out at the surface, but it is ordinarily concealed, especially in the eastern belt, by a relatively thin cover of barren clay. The barren overburden ranges up to

about 15 feet thick. The greatest depth to which barite-bearing clay has been found is more than 100 feet, at the Stephens mine. In most mines, barite lumps are scattered irregularly through the clay, but in some they are aligned along the vestigial bedding. In the eastern belt the ratio of clay and chert to barite (by weight) is about 7 to 1; in the central and western belts it is about 16 to1.

GEOLOGIC PROBLEMS

Although the Sweetwater barite deposits have been exploited for more than 80 years, they present many as yet unresolved geologic problems whose solution is important to further economic development of the area.

Origin and age.—It has generally been considered that the primary deposits are of hydrothermal origin and that the breccias in which they occur are of tectonic origin and of Appalachian age, the deposits thereby being late Paleozoic or post-Paleozoic in age. Most geologists would include these deposits in the "Mississippi Valley type."

Barite deposits occur at intervals along the strike of the Kingsport formation in the three belts, but primary deposits in bedrock are exposed only where residual deposits have been mined. They occur over a greater strike length in the central and western belts than in the eastern belt. Because of the limited exposures, it is difficult to ascertain the structural setting of the deposits. Dunlap (1945), by detailed mapping of exposures at the top of the Mascot dolomite and in the overlying Chickamauga limestone, demonstrated the presence of minor flexures trending northwestward, about normal to the regional strike, and as the axes of some of these flexures, projected from the exposed areas across the covered portion of the barite belts, passed through or near barite deposits, he concluded that they were responsible for the brecciation and thus controlled localization of the barite. This is the best interpretation of the structural control of the deposits that has been offered. These cross-folds are a ubiquitous feature of the outcrop belts of the Knox group, both in and beyond the Sweetwater district. It is the writer's opinion that they represent minor warping which took place at the close of Mascot time, when the Knox group was uplifted and subjected to weathering, before being submerged by the Chazy sea.

Minor angular unconformity between the Knox group and the overlying formations, though not detectable in any single exposure, was noted by the writer (unpublished TVA report) in a section constructed from closely-spaced drill holes at Douglas Dam, Sevier County, Tennessee, and by J. M. Kellberg (unpublished TVA report) in a similar section of an area near South Holston Dam in Sullivan County, Tennessee. Oder (1958, p. 53) mentions a similar situation in the Mascot-Jefferson City zinc district. Kendall (1958) has presented evidence which led him to conclude that some of the zinc deposits in the Kingsport formation at Jefferson City may have been deposited in breccias which are primary sedimentary structures. Thus, although direct evidence from the Sweetwater district is not available, there is a possibility that the minor flexures and the breccias are of Ordovician rather than late Paleozoic or post-Paleozoic age. Similarly, the barite deposits could be as old as Middle Ordovician. Unfortunately, the scanty exposures and the absence of underground

workings do not permit detailed study of the bedrock structures, and the answer to this problem may have to come chiefly from analogy with the zinc district.

There is general agreement that the residual deposits are the insoluble material left behind after removal of the soluble carbonates of the Knox group. The process by which the deposits were formed, however, was probably not a simple enrichment by removal of soluble matter; there was apparently solution, transportation and redeposition of the less soluble constituents, barite, fluorite and silica. One evidence of this is that both barite and chert occur in the residuum in much larger masses than have been seen in the bedrock. Adequate quantitative data are not available to establish the correct volumetric ratio between insoluble content of the bedrock and of the resulting residual deposits, but it appears that in many mines there may be more barite in the residuum than was available in the rock that was weathered to yield the residuum. Concentration by simple removal of soluble content should, especially in the areas of less than 15-degree dip, result in a wide scattering of barite in the residuum of the older, underlying formations. Yet in all mines which have been worked down to bedrock, residual deposits are concentrated directly above primary bedrock deposits, usually in the sub-outcrop of the lower member of the Kingsport formation, and a line drawn on the map through the stratigraphically lowest mines would coincide almost exactly with the trace of the Kingsport-Longview contact, except at places like the Stephens mine where there are prominent cross-folds. Unfortunately no bedrock is exposed at the deposits which are in residuum of the Longview and Chepultepec formations and it is not known whether they rest on primary deposits at the sub-outcrop, or were derived from deposits in the Kingsport formation which formerly overlay the area.

Removal of fluorite.—As a corollary of the situation just discussed is another problem, what became of the fluorite during weathering? Fluorite is a relatively insoluble mineral; in the Kentucky-Illinois fluorspar district residual fluorite, known as "gravel spar", has been an important commercial product, overlying the unweathered portions of the veins (Sutton, 1953, p. 143). Yet, in the Sweetwater district, fluorite, which is present in practically all exposed bedrock deposits, is not present in the shallower parts of the ore bodies. When fluorite begins to appear in the washer product, the mine operator recognizes it as a sign that the mine is approaching bedrock. Why did solution remove fluorite and not barite, and where did the fluorite go?

Relation to zinc deposits.—Deposits of barite and sphalerite occur in carbonate rocks—especially the Kingsport and Shady formations—throughout the Appalachian Valley, and although most deposits of one of these minerals contain the other, deposits containing commercial amounts of both are not known. The Eve Mills zinc deposit (Secrist, 1924, p. 130-133; Dunlap, 1947) which contains very little barite and fluorite, is directly along strike from and adjacent to the eastern belt, but seems to be entirely separate from it. The barite deposits of the Sweetwater district are closely similar to the Mascot-Jefferson City zinc deposits in structural and stratigraphic position. However, these two

types of mineralization do not appear to represent related zones of one mineral district (Laurence, 1939, p. 197-198), although that has been suggested (Behre, 1950, p. 38; Ohle, 1959, p. 773). Too little is known about the occurrence and distribution of the sulfide minerals in the Sweetwater district to permit any definite conclusions at this time about zoning within the district. Much more must be known about the occurrence and distribution of barite and fluorite beneath the residual deposits before their relation to the zinc deposits can be established. It is the writer's opinion that the two events—deposition of barite and fluorite at Sweetwater and deposition of sphalerite at Mascot-Jefferson City, may be entirely unrelated; the similarity results from the fact that the processes were similar and that both found favorable conditions in the same type of host rocks.

Physiographic position of barite deposits.—Barite mines occur at various elevations ranging from 760 feet above sea level near the Tennessee River at the northern end of the district and near the Hiwassee River at the southwestern end, to 1,100 feet in the middle portion of the district. They may be at or slightly above the alluviated bottoms of the Coosa level, or they may be only slightly below the upland (Harrisburg) level. However, many of the largest mines, including the Stephens which has produced at least half of the district's output, are in the middle part of the belt, near the divide between the Hiwassee and Tennessee Rivers and on or near divides from which drainage goes into three different tributaries. This suggests that the principal accumulation of barite was related to the development of the Harrisburg surface, and that the best place to find barite is on remnants of that surface.

The barren belt.—One of the more puzzling problems of the Sweetwater district is the apparent absence of barite in the barren belt. It may be noted that the same structural block contains the Mascot zinc deposits 45 miles to the northeast, and a small abandoned barite mine at Cleveland, Bradley County, 15 miles to the southwest (Swingle, 1959, p. 58-62), yet there are no mines or prospects in this belt within the Sweetwater district and, so far as the writer knows, not a single lump of barite has been turned up by the plow in this belt.

There are no obvious geologic or physiographic differences between the barren belt and the productive belts. Large areas of weathered Kingsport formation at favorable topographic elevations occur throughout the length of the belt. Neither the published cross sections (Rodgers, 1952, 1953, pl. 15) nor the alternative structural interpretations tried by the writer indicate any reason why the Kingsport formation of this structural block should not have been accessible to the barium-bearing solutions which mineralized the Kingsport formation of the other three blocks.

Examination of the soil map of McMinn County (Bacon and others, 1957) indicates one difference which may be significant. Soils overlying the Kingsport formation in the productive belts are classified chiefly with the Clarksville and Fullerton series, but those of the barren belt belong mainly to the Dewey (and, to a less extent, Decatur) series. All four of these series are considered to be of residual origin. The Clarksville and Fullerton are very cherty and are yellowish below the A_1 horizon. The Dewey and Decatur series are distinctly reddish; the Decatur is non-cherty and the Dewey contains only small chert

fragments in the deep subsoil (Bacon and others, 1957, p. 88-90). Examination of many exposures of these soils shows that the small chert fragments are stratified horizontally and that the reddish upper zones are actually old terrace deposits, probably related to tributaries rather than to the rivers. These old alluvial deposits are almost entirely restricted to the eastern and barren belts, though patches of them do occur farther to the west. They are composed almost entirely of material derived from Knox group residuum, though an occasional quartzite or vein quartz pebble can be found at most exposures.

Practically all the barite deposits in the eastern belt were covered by a noncherty red silty clay, 1 to 15 feet thick, described by many geologists as "colluvium" but properly identified by Gordon (1918, p. 58) as alluvium. The soil scientists, who are more concerned with utilization of soils than with their genesis, have mapped most of this clay with the residual soils. Superficially it looks much like the Dewey soils of the barren belt and also resembles red upland soils of the Blockhouse area in nearby Blount County, Tennessee, which were long called residual but recently were found by Neuman and Wilson (in press) to be ancient terrace deposits. There is some difference of opinion among the soil scientists, as expressed in oral discussions, as to whether all the soils mapped as Dewey are true residual soils. The writer suggests, as a possible explanation of the apparent absence of barite in the barren belt, that the widespread deep red soils represent ancient terraces which conceal the true residuum of the Kingsport formation and that barite deposits may actually underlie these soils, well below plow depths, in structurally and physiographically favorable areas. There is no obvious reason why the terrace deposits should be thicker and more extensive in the barren belt than in the eastern belt, but they appear to be so.

Silicification.—Residual chert which accompanies the barite deposits was derived from silica in the bedrock. It includes (1) primary, syngenetic chert, (2) epigenetic, (hydrothermal ?) chert, possibly related to the barite mineralization, and (3) silica derived from these two types but dissolved during weathering and redeposited as chert masses in the residuum. Most observers have commented on the enormous amount of chert which accompanies the barite, especially in the central and western belts. In part, this is undoubtedly due to the normally siliceous character of the formations of the Knox group, especially their western facies. One of the vexing problems yet to be solved in this district is the question of how much, if any, of the chert was originally deposited as epigenetic, hydrothermal silica along with the barite and fluorite deposits. As quartz does not occur in the mineralized veins, any such silica that is present must have replaced the country rock.

SUGGESTIONS FOR EXPLORATION

Most of the exploration for barite in the Sweetwater district has been by digging of testpits or churn drilling in areas where barite was accidentally turned up by the plow or bulldozer, in digging wells or cisterns, or by uprooted trees. As the area has been settled and farmed for more than 150 years, it seems reasonable to assume that most residual deposits occurring within plow

depth have already been discovered. Since, moreover, the bedrock deposits are not of economic value at present prices, exploration must be aimed at possible residual deposits which are below plow depth. Gravity surveys have been useful in Missouri (Uhley and Sharon, 1954) and should be tried in the Sweetwater district. A recent paper (Bloss, 1959, p. 37) suggests that geochemical prospecting of soils may also serve as a guide.

The writer suggests that the most favorable places for finding new barite deposits would contain two or more of the following features: (1) weathered lower member of the Kingsport formation, (2) a strong northwest-trending minor anticline crossing the regional strike, (3) location on or near a divide between two and preferably three major tributaries of the Hiwassee and Tennessee Rivers, (4) moderate elevation above the alluvial bottoms of the Coosa level, and (5) deep red soils of alluvial or colluvial origin.

This district presents a promising field for coordinated research in economic geology, geochemistry, geophysics, and geomorphology.

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THE UBIQUITOUS DIATOM—A BRIEF SURVEY OF THE PRESENT STATE OF KNOWLEDGE*

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ABSTRACT. Diatoms have distinctive and highly diversified tests or shells, similar to but not identical with opal in composition, accounting for their preservation as fossils in many kinds of sedimentary rocks and their importance to the geologist. The problem of unequivocally assigning diatoms and related minute unicellular or noncellular organisms to either the plant or the animal kingdom is increasingly being met by referring them to the Protista, which may be considered either as a separate kingdom of organisms or a confession of ignorance.

The earliest known large assemblage of marine diatoms occurs in rocks of Late Cretaceous age in California; the earliest known assemblage of nonmarine ones occurs in middle Oligocene rocks in Colorado. They are widespread over the globe, living today in virtually all non-toxic waters, occupying a wide variety of habitats. This wide distribution is explained by the ease with which these minute organisms may be transported over great distances in viable condition by means of water, wind, and attachment to larger organisms. The preservation of diatoms in sedimentary rocks is governed largely by the pH of the solutions penetrating the sediment, as the solubility of the diatom silica increases rapidly with increasing pH above 5. Even under such conditions diatoms may be preserved in calcareous concretions formed before the advent of the alkaline solutions.

The percentage of extinct species in any assemblage of diatoms increases with age of the enclosing sediment at rates comparable with those of mollusks, making them useful indicators of geologic time. The species in any assemblage still represented in living assemblages form the basis for paleoecological interpretations. The fact that large and representative assemblages of diatoms can be obtained from very small samples is of equal importance in both lines of investigation. The small size of the diatoms and their resulting ease of transportation introduces problems of reworking and contamination. The former is often difficult to evaluate; the latter can be virtually eliminated by systematic care and cleanliness in the collection and preparation of the samples.

INTRODUCTION

Diatoms are microscopic unicellular or noncellular organisms having a siliceous test or shell. At various times during the past hundred years they have been considered to be animals or plants, as they possess many of the attributes of both kingdoms. They are photosynthetic organisms possessing a combination of two dyes, diatomin and xanthophyll, which act in a manner similar to that of chlorophyll in higher plants in enabling them to synthesize complex carbon compounds from carbon dioxide and water through the action of sunlight. This constitutes one of their chief plant-like characteristics. On the other hand they store reserve food as lipoid fats and many have the power of independent movement, which constitute some of their animal-like characteristics. In view of these and other uncertainties, many students of these and other noncellular or unicellular organisms possessing this dual character have become uneasy about the arbitrary assignment of them to either the plant or animal kingdoms. Hogg (1860) early proposed the term Protoctista for various organisms not clearly classifiable as either plants or animals. Haeckel, (1866) proposed the term Protista for the same group, and his shorter name has remained and gained substantial following by the establishment of the journal Protistenkunde. Moore (1954) has recently reviewed the whole subject exhaustively, and has followed Hogg and Haeckel in recognizing a third kingdom of organisms, the Protista, for those microscopic unicellular creatures which for various reasons do not seem to fall happily into either the plant or animal kingdoms.

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Diatoms are composed of two siliceous valves, most of which fit together in a manner similar to the two halves of a flat pill box, with a siliceous band or girdle covering the junction. The exact composition of this siliceous valve is not reliably known, but appears to be a hydrous silica having the general formula of $SiO_2.nH_2O$, similar to the mineral opal. Although the refractive indices of the two forms of silica are very nearly the same, significant differences exist between the hydrous silica of opal and that of the diatom valve, probably related to the value of the integer *n*, which appears to be smaller for the diatoms than for opal. This is suggested by the greater resistance to corrosion by both strong acids and fairly strong bases characteristic of the diatom silica. Diatoms have long been considered to have "indestructible siliceous shells", but this apparent indestructibility is not sufficient to preserve them in porous sediments through which alkaline waters have moved.

The taxonomic classification of the diatoms has occupied the attention and exercised the imagination of many students of these organisms for more than a hundred years. At one time the classification was based upon the soft parts, in particular upon the arrangement of the chromatophores containing the photosynthetic dyes. This soon proved to be unworkable as these features were observable only in living diatoms, and the criteria subsequently accepted were based upon the shape, intricate structure, and often elaborate ornamentation of the siliceous valve and not at all upon the ephemeral soft parts. This is particularly advantageous for paleontological and paleoecological interpretations, as it virtually eliminates uncertainty in the generic and specific identifications made by biologists working with living diatoms and by paleontologists using fossil ones. Thus each dicipline can make effective use of the vast literature on diatoms, both living and fossil. The structure and ornamentation of the valve is remarkably constant for any given species, and is capable of precise biometric measurement, so that species can be identified with a high degree of certainty.

The classification of the diatoms was stabilized in 1896 by Schütt who proposed a satisfactory and workable one, which was published in Engler and Prantl's Naturliche Pflanzenfamilien. He considered the diatoms to be an order of Algae, Bacillariales (Diatomeae), divided into two suborders, Centrales, in which the structure and markings are more or less radial and related to a central point, and the Pennales, in which the structure and markings are related to a longitudinal line. A number of ambiguities in this classification developed during the nearly 40 years that it was almost universally adopted until Hendey (1937) proposed a modification of it, considering the diatoms as a Class of Algae, Bacillariophyceae, comprising one order, Bacillariales, which is divided into 10 suborders. This has proved to be a workable classification, eliminating many of the ambiguities inherent in the older system. The final step of resolving the doubt regarding their plant or animal status and assigning them to the Protista does no significant violence to Hendey's classification. The group comprises over 300 genera and between 12,000 and 16,000 species, depending upon whether one is a lumper or a splitter. Among this large number of species practically every known geometric shape has been used by these creatures, as well as many that bear no close resemblance to anything else.

Diatoms live today in marine, brackish, and fresh water, as well as in many more specialized habitats, such as hot springs, fresh water pools on oceanic ice floes, moist soil, peat bogs—in fact, wherever water non-toxic to them is present with at least partial exposure to sunlight. Some are pelagic, free-swimming organisms which spend their entire lifetime in the surface waters down to the depth to which actinic light can penetrate, usually not more than 60 to 100 meters. Others live attached to stones, aquatic plants and other submerged objects, or form the greenish brown slime coating the bottoms of shallow ponds and streams. Deep oceanic waters contain only the pelagic types, whereas the littoral zones contain both pelagic and benthonic forms. Estuaries and lagoons contain not only distinctly brackish types, but some truly marine forms and some truly freshwater forms which are introduced from the sea on one hand and from rivers or streams on the other. Similarly, deep nonmarine lakes contain only pelagic forms at some distance from shore, changing to mixtures of pelagic and attached forms in the shallower margins near shore. Near the mouths of streams, fluviatile forms, brought in by the streams mix with the lacustrine and marine types. Different species and in some cases genera of diatoms are confined to one or more of these habitats, reflecting their individual tolerances for water of various combinations of salinity, temperature, pH, nutriments, etc. Thus the study of fossil diatoms preserved in sedimentary rocks can yield paleoecological information of considerable value to the geologist, as defensible inferences can be made regarding the environment in which their enclosing sediments were deposited.

GEOLOGIC RANGE

Reported occurrences of diatoms in Paleozoic rocks have been found to be erroneous, as in each case more careful work has indicated that the presence of diatoms was due to contamination with younger material, either in the laboratory or in the field. The whole subject of the existence of pre-Mesozoic diatoms was extensively explored by Pia (1931), who found no authentic occurrences. Although a few diatoms have been found in what appear to be Jurassic rocks, the earliest large assemblage of well preserved ones have come from the marine Moreno shale of Late Cretaceous age in the Panoche Hills, Fresno County, California. Hanna (1927, 1934) described many new species and some new genera from this locality, to be followed by Lefébure and Chenevière (1939), and by Long, Fuge, and Smith (1946), who added considerably to the original list so that today the Moreno assemblage consists of 36 genera and 120 species. This assemblage contains an overwhelming proportion of extinct species, and is so diversified that it is highly probable that the origin of the diatoms as a biologic unit may have taken place much earlier, either in the earlier Mesozoic or possibly even sometime during the Paleozoic. Marine diatoms are known from rocks representing all succeeding units of geologic time up to and including the Recent, with increasing diversification until today there are hundreds of described genera and many thousands of species.

Apparently the nonmarine diatoms have had a somewhat shorter geologic history, the earliest one known having been found in the Florissant lake beds

of middle Oligocene age near the town of Florissant, Teller County, Colorado Here again, the degree of diversification of the Florissant assemblage, although somewhat less than in the case of the marine Moreno shale assemblage, is great enough to suggest strongly a pre-Oligocene origin for the non-marine diatoms. The known geologic record of the nonmarine diatoms is both shorter and less continuous than that of the marine ones, as the next earliest occurrence is in the lower Virgin Valley beds of Merriam (1907). These beds in Humboldt County, Nevada, have been called late middle Miocene in age on the basis of extensive vertebrate faunas. Diatoms are represented in all succeeding units of geologic time.

The sudden appearance of large numbers of individuals and species of marine diatoms in Late Cretaceous time and the equally sudden appearance of the nonmarine forms in middle Oligocene time poses a question which has not been answered satisfactorily. Their absence in earlier rocks is due, most probably, to lack of preservation. In this connection, Roy A. Bailey, of the U. S. Geological Survey (oral communication, December, 1959) has stated that no preserved opals are known from pre-Cretaceous rocks. Although the diatom silica appears to be more stable than opal in the laboratory, it is possible that its stability in the rocks is not enough greater than that of opal to be preserved.

GEOGRAPHIC DISTRIBUTION

Diatoms today inhabit the photic zone, usually the upper 60 to 100 meters, of all marine waters of the globe in such quantities that they constitute a major item of food for larger animals, and have been called the "grass of the sea". Their concentration in sea water ranges from a few thousands of diatoms per liter to over 10 million per liter. The oceanic, pelagic types are carried for long distances by ocean currents, on the feet and feathers of birds, by fish, and by the wind. Although many species prosper better in somewhat restricted habitats governed by optimum salinity, temperature, pH, and the like, the random scattering of these organisms by the agencies briefly outlined results in many of them landing in a satisfactory habitat and hence quickly establishing themselves far from their point of origin. As will be shown later this rapid transportation over wide areas has also been a factor in the distribution of these organisms in the geologic past.

The nonmarine diatoms have a similar wide distribution and today different assemblages inhabit virtually all types of aqueous environments, lakes, rivers, brackish estuaries, ponds, bogs, marshes, ice floes, hot and cold springs, and even moist soil. In fact, it may be said safely that diatoms are the most ubiquitous creatures that leave hard parts for the Cenozoic fossil record. This wide distribution is made possible by the fact that diatoms are able to remain viable during long periods of almost complete desiccation, and some are able to produce heavy-walled virtually hermetically sealed resting spores which remain viable even after long periods of complete desiccation.

TYPES OF ENCLOSING SEDIMENTS

Fossil diatoms have been found in all types of water laid sediments with the exception of very clean sandstones and conglomerates. They may occur in

such great numbers that they form a diatomite, a micro-coquina made up almost entirely of diatoms, in which very small amounts of clay or volcanic ash may be present as impurities. They may occur as a major or minor constituent in shales, claystones, mudstones, siltstones, clayey or silty sandstones, and in both bedded and concretionary limestones. They also commonly occur in water-laid tuffs. Inasmuch as they are nearly always present in all bodies of non-toxic water, the proportion of diatoms to clastic materials in a sediment often provides a useful clue to rapidity of deposition. Pure diatomites, which are deposited very slowly, for example, clearly indicate a period during which little or no clastic material was being carried into the basin of deposition. The world-wide correlation between thick deposits of diatomite and contemporary vulcanism has been observed by many workers, beginning with Ehrenberg (1846), and has been ably summarized by Taliaferro (1933). Volcanic ash constitutes one of the most readily available sources of soluble silica needed by the diatoms, and shards of volcanic glass are often found in beds of diatomite, particularly when they attain thicknesses of more than a few feet. Reinhold (1937) has considered that the common association of diatoms and volcanic ash may be due in part, at least, to the fact that the diatom test would be less likely to be dissolved in water saturated with silica derived from the volcanic ash.

PRESERVATION

The known distribution of diatoms in sedimentary rocks is probably more closely dependent upon their preservation than upon their original occurrence in the different basins of deposition. The absence of these ubiquitous creatures in a sedimentary rock commonly requires a more searching explanation than does their presence. The most obvious explanation and one for which there is abundant field evidence, is that the diatoms were present in the sediments at one time, but have been leached by alkaline or silica starved water either immediately after deposition or by percolating alkaline ground water at some time after deposition or even following tectonic movements which raised the beds above sea level. Diatom silica is normally fairly stable and can resist leaching for long periods of geologic time as long as the pH of the sediment does not exceed 4 to 5. The solubility of silica (SiO_2) increases five fold between pH 5 and 8. No data are available for the solubility of diatom silica ($SiO_2.nH_2O$), but it must be much greater than the non-hydrated form. Therefore the finding of an abundant assemblage of well preserved diatoms in the Moreno shale of Late Cretaceous and Paleocene (?) age suggests that the pH of any water in contact with the formation was always quite low, probably never much above pH 4 or 5. Even when the pH has been higher and hence leaching has occurred, it is often possible to find well preserved diatoms in calcareous concretions, if the calcium carbonate forming the concretions came into the sediment before any alkaline waters. At first sight this appears to be a large *if*, but the following examples will illustrate the fact that this has happened in at least some instances.

One of the best examples of the progressive effects of leaching is the classic and extraordinarily well exposed section of the Miocene Monterey formation in Chico-Martinez Creek on the east slope of the Temblor Range, northwest of the

town of McKittrick in Kern County, California. Here the upper 800 feet (stratigraphically) of the section consists of punky diatomite composed almost entirely of well preserved diatoms. Stratigraphically below this unit is one comprised of over 1000 feet of porcellanite, massive in the upper part and laminated in the lower part. This unit contains no preserved diatoms in the porcellanite although numerous impressions of them can be seen on fresh bedding surfaces with a hand lens, indicating that they were once present in abundance. Stratigraphically below this is a unit of over 4000 feet of laminated porcellanite (or porcellaneous shale) in which not even impressions of diatoms can be found, although the rock is otherwise similar to the unit above it. No diatoms can be recovered from any of the porcellaneous rocks below the 800 foot diatomite at the top of the section, because the diatoms have been completely dissolved and their silica redeposited almost *in situ*, in fact this is almost certainly the process whereby the rocks in the lower part of the section became porcellaneous. Stratigraphically below the laminated porcellanite the proportion of cherty shale and chert increases through another interval of over 1000 feet, and here, of course, no diatoms can be recovered.

At various stratigraphic levels throughout the porcellanites and cherts below the 800 foot diatomite at the top of the section, zones of calcareous concretions occur. Some of these are discontinuous beds of nodules or ovoid concretions, ranging from about an inch to over a foot in major diameter, and some are fairly continuous calcareous beds of uneven thicknesses up to a maximum of a foot or more, traceable for thousands of feet along the strike. These calcareous concretions yield abundant and well preserved diatoms from their unweathered interiors. Bramlette (1946), who studied the Monterey formation in great detail, has traced fine laminations in the beds of barren porcellanite into and through some of the diatomaceous calcareous concretions, indicating that the calcium carbonate was secondary, but earlier than the alkaline solutions which leached the diatoms. Thus it appears certain that the Monterey formation, having in many areas in the Coast Ranges a stratigraphic thickness of approximately 1 mile, was once largely composed of diatoms, most of which have been leached and the silica from their tests redeposited *in situ* to form porcellaneous shale and chert. The diatoms are clearly rock builders of gigantic proportions. The number of diatoms involved in building such a body of rock becomes a staggering figure in view of the fact that actual counts of diatoms from the Monterey formation have resulted in a realistic figure of about 21,000,000 diatoms per cubic inch.

Another example, this time from nonmarine rocks, will serve to illustrate another aspect of the preservation problem. The Furnace Creek formation of Miocene or Pliocene age in Furnace Creek Wash on the east drainage into Death Valley, Inyo County, California consists dominantly of lacustrine claystones and mudstones containing in places great quantities of sodium and calcium borates. In fact, before 1926 the largest commercial production of borates came from the Furnace Creek formation. As a result, the pH of these sediments is high, up to pH 9 or over in many places. Intensive search has revealed no diatoms in the claystones and mudstones of this formation, but a few have been

found in interbeds of freshwater limestone. These limestones are rarely concretionary, but are more commonly bedded, with thicknesses up to several feet. In one instance, a clean white limestone at the top of the Furnace Creek formation yielded an abundant assemblage of well preserved diatoms of late Pliocene or early Pleistocene age. In the lower part of the same formation, the diatoms recovered from limestones are often so badly leached that specific identification is difficult. However, in the same beds that contain the leached ones are several species of diatoms whose Recent counterparts live in hot springs, some of which are still active nearby. The hot springs diatoms, of which the most common in the Furnace Creek formation is *Denticula thermalis* Kützing, are always well preserved, indicating that the silica forming their shells is less soluble in alkalies than that found in other diatoms. It appears probable that this may be due to somewhat less hydration of the silica.

USE IN STRATIGRAPHY

The diatoms, both marine and nonmarine, like most other groups of fossils, have both short ranging and long ranging species. Each assemblage of diatoms, with the possible exception of those from very late Pleistocene rocks, will have some extinct species as well as some that are still represented in living assemblages elsewhere.

The following table lists the percentage of living species present in a number of marine diatomaceous rocks from various places in the world, and from Late Cretaceous to late Pleistocene in age. To give some comparison with the marine mollusks, Lyell's early, 1832, figures are given, followed by Schuchert's, 1915, revision:

PERCENTAGE OF LIVING SPECIES OF MARINE DIATOMS AND MOLLUSKS

Mollusks		Age	Diatoms	
Lyell 1832	Schuchert 1915			
90-95	95-100	Pleistocene	Atlantic Coast	95-100
35-50	50-90	Late Pliocene Late Miocene to Middle Pliocene*	San Joaquin fm., Calif. Sisquoc fm., Calif.	85 70
17	20-40	Late Miocene Middle Miocene Early Miocene	Upper part of Monterey fm., Calif. Calvert fm., Md. & Va. Upper part of Temblor fm., Calif. Santos shale, Calif.	60 44 36
	10-15	Oligocene Eocene and Oligocene*	Oamaru, New Zealand Kreyenhagen fm., Calif.	21 24
3.5	1.5	Eocene	Jutland Barbadoes Ananino, Simbirsk, Russia	16 12 11
		Late Cretaceous and Paleocene (?)	Moreno shale, Calif.	5

* Age of rocks from which diatoms were obtained.

The wide distribution of diatom species and assemblages makes them particularly useful for stratigraphic correlation. For example, the Temblor formation of Miocene age in Oil Canyon, north of Coalinga, California, contains a diatom assemblage composed of over 50 percent of extinct species as well as number of extinct genera. Many of these species and some of the genera are known in California only from the Temblor formation, although the diatoms from both older and younger formations nearby have been studied and compared. In Maryland, the Calvert formation of middle Miocene age contains the same extinct species and genera of diatoms, which occur both in outcrops forming cliffs along the west side of Chesapeake Bay from Fairhaven to south of Plum Point, along the Patuxent River near Nottingham (Ehrenberg, 1854; Boyer, 1904) and at depths of 1000 to 1140 feet in cores from a dry hole drilled by the Ohio Oil Company near Salisbury, Maryland (Lohman, 1948). No older diatomaceous strata belonging to the Choptank and St. Marys formations contain different diatom assemblages. Both the Temblor and Calvert formations have molluscan faunal elements which were independently determined as middle Miocene in age. The two diatom assemblages from the east and west coasts are so strikingly similar that they could easily be confused. Other examples could be cited, but there is no need to labor the point.

Another advantage enjoyed by the student of diatoms is the fact that a large and diversified assemblage of fossils can be obtained from a very small sample of rock. This permits accurate frequency counts to be made of the different species in an assemblage, which often permits sharper correlations over relatively short distances.

The nonmarine diatoms present more problems to the stratigrapher than do the marine ones on account of the much wider range of habitats they occupy. It is quite possible, for example, for two lake basins of identical age to contain quite different diatom assemblages. The many factors which control the growth of different species of diatoms, such as temperature, salinity, silica, and pH of the water, to mention only the most obvious ones, vary over much wider limits in nonmarine water bodies than in the sea. On the other hand, the dependence of most species upon optimum or at least tolerable values of these factors, make it possible to infer such optima from a fossil diatom assemblage and gain much paleoecological information of value to the geologist in reconstructing the environment in which the particular sedimentary rock was deposited. The precision with which this can be done would be greatly increased if all students of living diatoms would make more systematic and quantitative measurements of these essential factors, at the same time that the collections of living diatoms were made. It might result in fewer collections, but these would have vastly greater significance. The paleoecologist is at best an extrapolator and he is greatly hampered if he has only meager generalities from which to extrapolate.

Within the limitations set by these paleoecological considerations, the nonmarine diatoms are equally useful in age determination and for the correlation of strata. Fortunately, as new areas are studied, many of the lake sediments, fall into well defined patterns of paleoecology with the result that an increasing

amount of confidence can be placed upon age determinations based on nonmarine diatoms. The percentage of living species in any nonmarine assemblage decreases with age at about the same rate as the marine counterparts.

REWORKING AND CONTAMINATION

The ease with which diatoms can be transported, although insuring wide and rapid dispersal, is also something of a disadvantage, as it makes reworking from other beds and contamination from other collections very easy. In general, this problem can be divided into two parts having distinctly different origins, but which affect the end product—a slide of diatoms under the microscope—in much the same way.

Reworking is obviously related to the normal geologic processes of erosion, transportation, and deposition. Fossiliferous rocks are constantly being eroded and the fossils in them transported to new sites of deposition, where they mix with Recent organisms. In the case of the larger fossils the transported distances are usually not great and the fossils commonly suffer obvious abrasion in the process. Diatoms, on the other hand, again by virtue of their minute size, can be transported great distances not only by water, but by wind, with little evidence of abrasion. For example, the Calvert Cliffs along the southwestern shore of Chesapeake Bay contain beds of both mollusks and diatoms of middle Miocene age. These are constantly being eroded and the product acted upon by each tide and by waves. The mollusks eventually move to the sandy beach, where they are abraded and often completely ground up by waves and tides so that it is rare for them to be redeposited in the bay sediments in anything like their original condition. The diatoms, on the other hand, remain in suspension in the turbulent waters of the shore and are carried out to the deeper parts of the bay and deposited. It is quite probable that many of them are carried out to the open sea and deposited on the continental shelf. The abrasion on the diatoms in this process is very slight as they are protected by being surrounded by water. Thus middle Miocene diatoms are being deposited in Recent sediments somewhere in the bay or along the continental shelf. Fortunately, the actual number of older diatoms being incorporated into sediments in this manner is usually small compared with the vastly greater number of Recent forms with which they become mixed. The dilution factor is so large, that in any assemblage of Recent sediments from the bottom of the bay, the Calvert forms would be reported as very rare if found at all. Nevertheless occasions can arise where the dilution with Recent forms is not so great, and constant vigilance is needed in interpreting such assemblages.

In some instances, however, diatoms which have been transported for long distances constitute a large proportion of the diatom assemblages found. For example, in a study made of a series of deep sea cores collected across the North Atlantic between Newfoundland and Ireland (Lohman, 1941) it was found that the diatoms from the cores taken at depths of water ranging from 3,230 to 4,820 meters were not the cold water forms now inhabiting the surface waters of the North Atlantic at that latitude, but contained a high proportion of warm water forms which live today in the warmer equatorial waters.

The most probable explanation is that the diatoms, after death, started to sink somewhere near the equator and were carried northward to their final resting place at 50° N. latitude by the northeastward flowing Gulf Stream. It was estimated that as much as several hundred years may have been required for this trip. Fortunately, most diatomaceous sediments encountered by the micropaleontologist are not deposited under such a combination of extremely deep water and a persistent ocean current. This study, however, emphasizes the need for caution in any paleoecological interpretations made on the basis of diatoms collected from the bottoms of deep ocean basins.

An obvious case of the reworking of older fossils into younger sediments occurred in the examination of a series of drill cores from twelve localities widely scattered along the west side of the San Joaquin Valley, California (Lohman, 1954), all of which penetrated, at depths of from 250 to 600 feet, a lacustrine deposit called the Corcoran clay by Frink and Kues (1954). All of these cores yielded a large assemblage of dominantly freshwater diatoms of late Pliocene to early Pleistocene age. One of the cores also yielded a few individuals of several extinct species of marine diatoms, two of which are known only from rocks of Oligocene age, one with a known range of Late Cretaceous to middle Miocene, and several with longer geologic ranges. At the present time many square miles of diatomaceous strata of the marine Kreyenhagen formation, known to contain these species, are exposed to erosion in the hills bordering parts of the west side of the valley. It appears quite probable that similar exposures of the Kreyenhagen formation were available to erosion during the deposition of the Corcoran clay, and that a few of these were carried into the basin. In working with such minute organisms as diatoms, one must be constantly alert to the possibilities of reworking, which are usually less obvious than the case just described.

The problem of contamination of one collection with another is an equally vexing one, but one which can be practically eliminated by scrupulous care and cleanliness both in making collections in the field and in preparing them for study in the laboratory. It is virtually certain that most of the reported occurrences of diatoms in rocks of Paleozoic age were the result of such contamination.

A convenient and safe container for individual collections of diatomaceous rocks in the field is the Kraft paper sample bag, about 4 x 7 inches in size, or even smaller. These are fitted with a metal strip which can be used to effectively seal the bag after the top has been turned down for a few tight folds. In the laboratory only simple glassware, such as beakers, capable of being effectively cleaned should be used, and collections undergoing the various chemical and physical operations should never be allowed to become dry. They should be kept covered with glass or plastic bell jars between operations. The laboratory preparation of the collections is essentially a problem of disintegrating the rock into its constituent particles without losing or breaking the diatoms, and then concentrating the diatoms by the removal of all other rock constituents. Such rigid requirements cannot always be met, nor is it always economically de-

sirable, as the removal of the last fraction of fine clastic material can be very difficult. A brief resume of the preparation process, which has since been improved in many respects, is given in my paper on the North Atlantic cores (Lohman, 1941, p. 56).

THE PUBLISHED RECORD

To list even a fraction of the thousands of papers and monographs on diatoms that have been published during the past 140 years would require much more space than is available. The quality of these papers is variable, reflecting both the competence of the men who wrote them and the quality of the often primitive optical equipment at their disposal. With a few notable exceptions, few of these men were either geologists, paleontologists, or stratigraphers, with the result that the present day micropaleontologist finds many of these monographs excellent for the identification of species but woefully lacking in vital stratigraphic data. Many authors, particularly the earlier ones, divided all diatoms into Recent and fossil, without regard to the relative ages or geologic ranges of the fossil forms. Some of the papers on living diatoms contain ecological data of value to the paleoecologist, others none. Many of these papers, particularly the older ones, are available only in the libraries of the larger universities and scientific institutions, and in the personal libraries of specialists. In spite of all these obstacles, the ubiquitous diatom is here to stay and promises to become an increasingly useful tool in stratigraphy and paleoecology.

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POSSIBLE EXPLANATION OF DIVERSE STRUCTURAL PATTERNS IN SOUTHERN NEVADA*

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ABSTRACT. In southern Nevada a belt of folds and thrusts striking generally north is transected by a northwest-trending fracture zone on which right-lateral movement has occurred. This movement was in progress while some of the thrusts and folds were active. A suggested explanation assumes that the two structural patterns reflect independent controls at different levels in the crust.

INTRODUCTION

Close association of two or more distinctive patterns of faults and folds may suggest superposed effects of successive deformations. Study of critical features in an area with intersecting trends commonly yields clues for distinguishing older and younger effects, giving a basis for recognizing stages in the tectonic history of the area.

A wide belt west and north of Las Vegas, Nevada (fig. 1), in which major

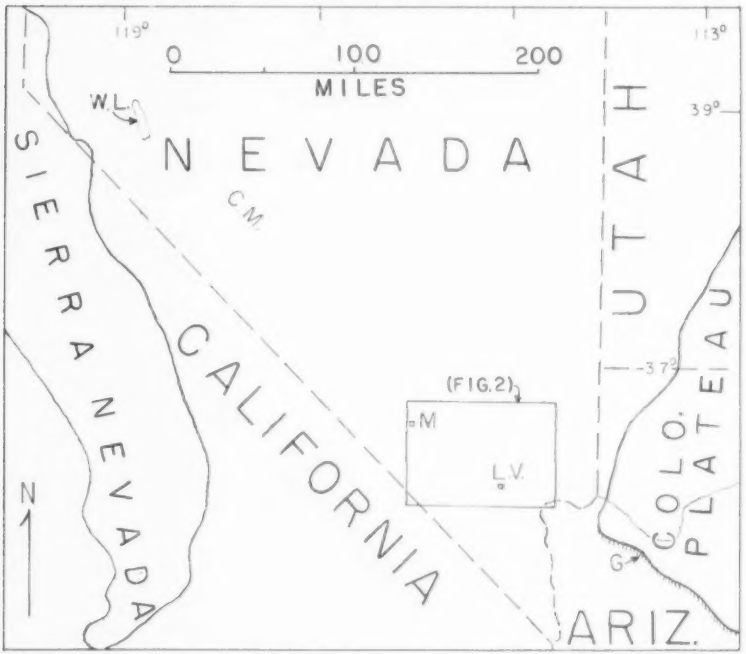

Fig. 1. Index map. W. L., Walker Lake; C. M., Cedar Mountain; M, Mercury; L. V., Las Vegas; G, Grand Wash Cliffs. Small rectangle encloses area of fig. 2.

thrust faults and associated folds trend generally northward, is transected by a zone of right-lateral displacement that trends northwestward. Movement along

* Publication authorized by the Director, U. S. Geological Survey.

this zone continued after the thrusting ended; but impressive evidence suggests that the two types of displacement were in part concurrent. As this possibility may seem anomalous from some points of view, analysis of the evidence is in order.

GENERAL FEATURES OF THE AREA

Topography.—Las Vegas Valley, which occupies a structural trough several miles wide, is flanked on both sides by high ranges (fig. 2). Several ranges that trend generally north, in accord with major structural elements, end abruptly southward at the north margin of the valley. South and west of the valley is the Spring Mountain Range, the dominant topographic unit of the area. The crest of this range trends northwest, generally parallel to the valley; but in much of its course the topographic axis is strongly discordant to the dominant structural grain, which varies in direction from north to northeast. Ridges controlled by the strike of beds extend consistently northeast in a belt 10 to 12 miles wide between the main range and Las Vegas Valley.

Kinds of rock.—Bedrock in the ranges consists almost entirely of sedimentary formations that were deposited in the eastern part of the Cordilleran geosyncline. The Paleozoic section, thickening gradually westward from the Colorado Plateau, totals about 8,000 feet in Frenchman Mountain, where all systems except the Silurian are recognized. Rocks of this system also appear as the section thickens northwestward at accelerating rate. In a section through the Spring Mountain Range extending 40 to 50 miles northwest from Las Vegas, the maximum thickness of Paleozoic strata, as measured upward from the lowest horizon known to contain Cambrian fossils, exceeds 22,000 feet. An older conformable section of beds which have yielded no significant fossils makes the total thickness, from the top of the Permian system downward, more than 30,000 feet. A comparable section is exposed north of Las Vegas Valley. Triassic and Jurassic formations, preserved only in a limited area west of Las Vegas, bring the known total thickness of sedimentary beds laid down before the onset of late Mesozoic orogeny to about 35,000 feet.

The oldest formations exposed in the area, including those firmly dated as Early Cambrian, consist chiefly of clastic sediments (Nolan, 1929). Above these is a much thicker sequence, ranging in age fgrom Middle Cambrian to Permian, made up largely of limestone and dolomite. Clastic materials predominate in the Mesozoic formations. The Aztec sandstone of Jurassic (?) age, more than 2000 feet thick, is made up of nearly pure quartz sand in thick beds with large-scale cross lamination. Normally red-brown from ferric oxide in a weak cement, much of the sandstone in this area has become pale yellow to white because of leaching by circulating solutions in fault zones.

STRUCTURE

Thrust faults.—The earliest record of orogeny is presumably post-Jurassic, as the youngest rock formation involved in this area is the Aztec sandstone. Probably the deformation occurred near the middle of the Cretaceous period; orogenic deposits of that date are a prominent feature of the Muddy Mountain thrust zone about 40 miles northeast of Las Vegas (Longwell, 1949), and this thrust appears to be a good sample of the regional pattern. In the Keystone

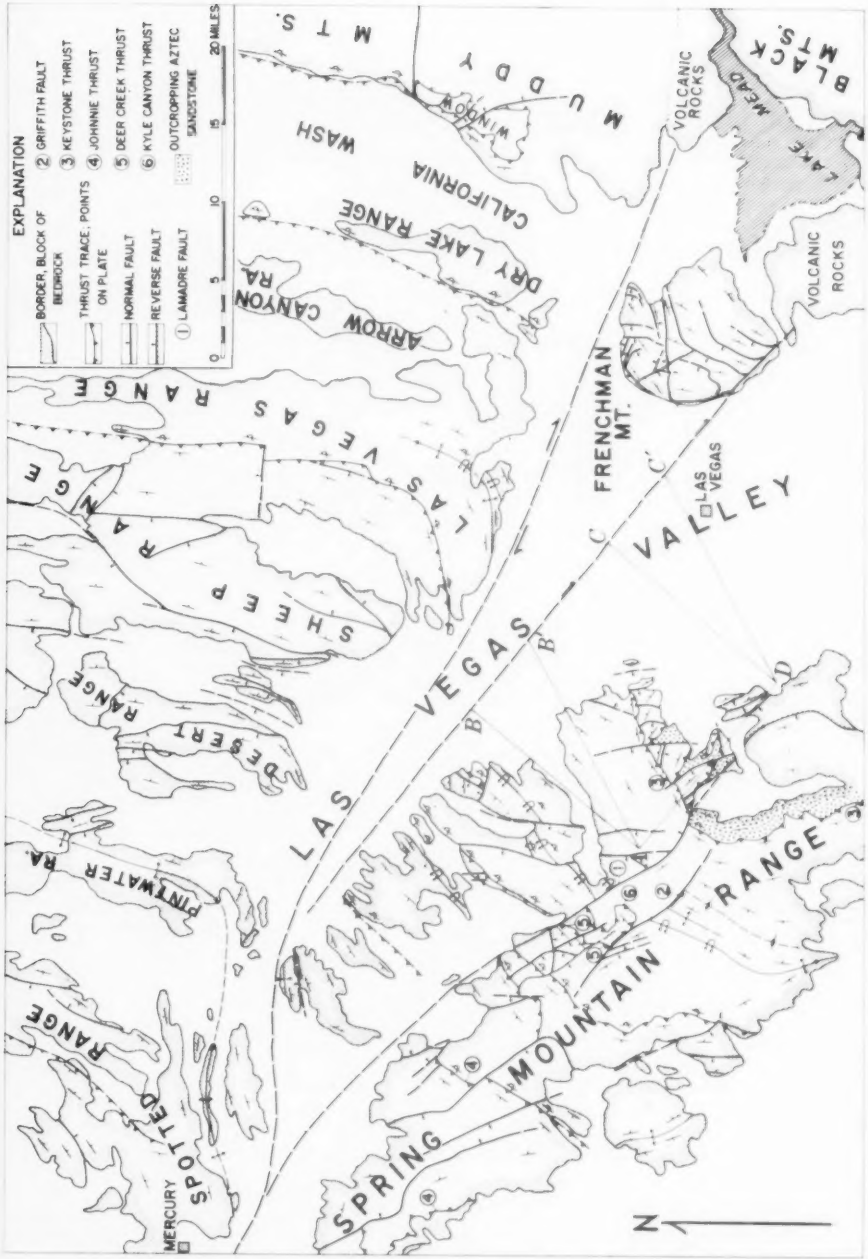

Fig. 2. Tectonic map of Las Vegas Valley and adjacent areas. Special symbols are explained on the figure. Standard symbols for strike and dip, anticlinal and synclinal axes, overturned anticline.

thrust, at the southeast end of the Spring Mountain Range, Middle Cambrian dolomite has overridden the Aztec sandstone. Thrusts farther northwest, with traces essentially parallel to that of the Keystone thrust, have been much eroded and are now visibly associated with no formations younger than late Paleozoic. So far as can be judged from exposures in a wide region, the thrusts are confined to the sedimentary section. In the Johnnie thrust, well exposed in the northwestern part of the Spring Mountain Range (Nolan, 1929), a plate of which the strong Stirling quartzite of Early Cambrian age forms the base moved eastward over the older Johnnie formation. This movement was made possible by buckling and shearing across the entire plate, the base of which then rode up across younger formations (Longwell, 1945, p. 422). The Keystone thrust plate, which has in its basal part Middle Cambrian carbonate beds that normally rest on the Pioche shale of Early and Middle Cambrian age, probably was sheared over that weak unit, along a "bedding thrust", west of its visible transgression of younger formations.

Overturned anticlines that parallel thrust traces in the Spring Mountain Range are, in all probability, incipient stages of thrusts like the Johnnie and Keystone thrusts. Several anticlines of this kind appear essentially as offset continuations of thrusts whose traces end abruptly against strike-slip fault zones that accommodated adjustments in compressive deformation (fig. 2).

One feature of the Keystone thrust commands exceptional interest. From a strike that averages nearly north through 15 miles or more, the trace of the thrust bends to a direction about N. 60° E. which is maintained through nearly 10 miles to the southwest margin of Las Vegas Valley (figs. 2, 3). This change in strike conforms to a pattern common in the Spring Mountain Range. Unique in the Keystone thrust is the change, near the bend in strike, from simple to complex structure below the thrust contact. South of the bend the Aztec sandstone is in normal stratigraphic sequence with beds beneath it and strikes generally parallel to the thrust surface. From the bend northeastward the hanging wall rests on a succession of fault blocks that are tilted strongly toward the northeast along normal faults striking nearly north (fig. 3). In four of these blocks Cambrian beds are in thrust contact with the Aztec sandstone, just as in the main thrust south of the bend. Although this relation is not visible in the tilted blocks farthest east in the series, presumably this is so only because the tilting and erosion of these blocks are insufficient to expose the thrust contacts.

Close study of the relations outlined above leads to a favored interpretation, as follows: While the Keystone thrust was active it was transected by a set of north-trending fractures induced by localized strains. With continued movement of the thrust plate the blocks bounded by these fractures were tilted and overridden. Origin of the strains responsible for the fracture pattern must be sought in a broader structural analysis.

Las Vegas Valley shear zone.—Extensive right-lateral displacement along Las Vegas Valley is established convincingly by several lines of evidence. Fold axes, traces of thrusts, and strikes of tilted beds, in approaching the valley along north-south lines bend strongly eastward south of the valley, strongly westward north of it (fig. 2). The Frenchman Mountain block, tilted steeply

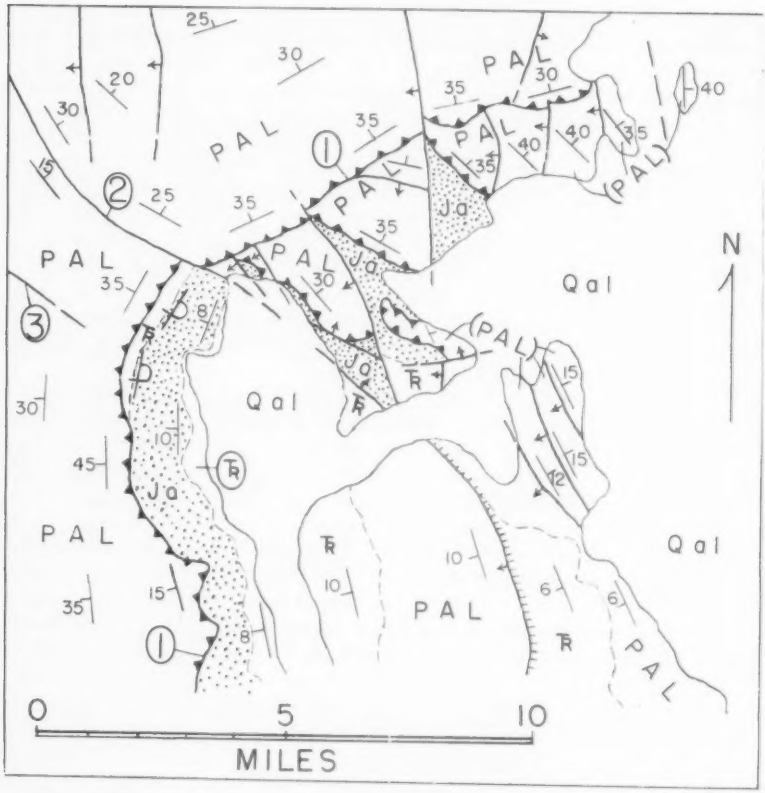

Fig. 3. Details of Keystone thrust and related features. (1), Keystone thrust; (2), La Madre fault; (3), Griffith fault; PAL, Paleozoic formations; $\overline{\text{TR}}$, Triassic formations; Ja, Aztec sandstone (Jurassic ?); Qal, Quaternary alluvium. Other symbols as on fig. 2.

to the east, has the strike of beds in its northern part turned sharply eastward; several fault slices crossed in approaching the northern margin are offset progressively eastward; and features along the south margin of the block are displaced to the east in relation to fault blocks south of it. Stratigraphic evidence suggests lateral displacement of large measure. In the Frenchman block the Ordovician system is represented by dolomite beds, 100 feet or less in total thickness. Six miles north from the northernmost exposure of this unit, across a belt of alluvium that masks the shear zone, incomplete outcrops above a thrust reveal three distinctive formations of Ordovician age, including the distinctive Eureka quartzite, with total thickness at least 800 feet. These formations extend continuously to the north-northwest. A thrust plate in the southern part of the Las Vegas Range reveals a thick characteristic section of the Wood Canyon formation and Stirling quartzite below the Pioche shale; only 16 miles

southeast, at the front of Frenchman Mountain, a thinner section of the Pioche is separated from Precambrian gneiss and schist by only 150 feet of sandstone, closely similar to the Tapeats sandstone of Early Cambrian age in the Grand Canyon. Devonian and Mississippian formations in the south end of the Sheep Range have members made largely of clastic materials; whereas directly across Las Vegas Valley, in ridges that branch from the Spring Mountain Range, the two systems are represented almost entirely by carbonate rocks.

Thus the available evidence indicates large lateral displacement on the shear zone, but it gives no quantitative measure. We may speculate that the Muddy Mountain thrust, the easternmost large thrust fault north of the shear zone, was originally the northward continuation of the Keystone thrust which has an analogous position in the belt of thrusts on the south. This would involve lateral movement of 25 miles or more (fig. 2). Displacement of such measure seems reasonable as judged by the qualitative, particularly the stratigraphic, evidence. Attempts to match given folds and thrusts north and south of the shear zone, however, involve assumption that these structures were developed continuously across the zone and offset by later horizontal movement. Another possibility is illustrated in the Jura chain of Switzerland and France, where patterns of strong folds were developed independently on opposite sides of concurrent strike-slip faults (Heim, 1919, pl. xx). If the Las Vegas Valley shear zone was active during the late Mesozoic orogeny, structural patterns north and south of the zone may have differed from the start, and such features as the Muddy Mountain and Keystone thrusts may have ended along this zone at their inception. Evidence on relative dates of the two structural patterns must be examined critically.

Inception of the shear zone.—The strongly tilted stratigraphic section in the Frenchman Mountain block includes the Horse Spring formation of Early Tertiary or Late Cretaceous age which was laid down after most if not all the thrusting movements in the Muddy Mountain belt (Longwell, 1928, p. 121). As tilting of the Frenchman block logically is related to displacements in the shear zone, there was important movement in this zone after the main orogeny ended. Possibly the zone was active through much of Tertiary time, as basin deposits of the Muddy Creek formation, probably of Pliocene age, are steeply tilted near and parallel to the shear zone southwest of the Muddy Mountains. Major strike-dip faults commonly have had long histories; the San Andreas fault zone, still active, operated during much of Tertiary time and possibly dates from the Mesozoic era (Hill and Dibblee, 1953, p. 449). Important in the present analysis is evidence that lateral displacement along Las Vegas Valley was in operation during the main orogeny as well as later.

Strongly suggestive evidence appears in connection with the nearly parallel La Madre and Griffith faults, which extend northwest from the area of flexure in the trace of the Keystone thrust (fig. 2). Movement on both faults has been in large part horizontal, in a direction essentially parallel to the Las Vegas Valley shear zone. The longer of the two, the La Madre fault, is clearly identified through more than 15 miles, and probably continues much farther, under alluvial cover, as part of a major fault zone near the northeast base of the high Spring Mountain block. As shown in figure 2, the logical course of the

La Madre fault forms nearly a straight continuation of the main shear zone in the vicinity of Mercury, whereas the belt along Las Vegas Valley, indicated as the zone of largest displacement, has a curving course. This pattern of echelon branching, common in major zones of lateral movement, is well illustrated in detailed maps of the San Andreas fault zone.

Along exposed parts of the La Madre fault the bedrock has been ground to small-size rubble in a zone tens of feet wide. Functioning of the fault in orogenic adjustment is convincingly established by abrupt ending of the Kyle Canyon thrust at the fault, northeast of which an overturned anticline, in offset position, continues the belt of compressive deformation with northeast strike (fig. 2). The Kyle Canyon thrust, in which Mississippian and Devonian formations override Permian, and the Deer Creek thrust, bringing Cambrian over Pennsylvanian, end abruptly southwestward at the Griffith fault. Shortening by thrusting northeast of the Griffith fault is essentially equalled by folding to the southwest of it. Thus some characteristic segments of the orogenic belt are gentically related to strike-slip faults with northwest strike, one of which—the La Madre fault—may be a branch of the shear zone in Las Vegas Valley. The relations outlined above suggest that folding and thrusting in the belt between the La Madre fault and Las Vegas Valley were in progress while activity in the shear zone was causing lateral dislocation and change in strike.

The tilted fault blocks, fragments of a thrust overridden by the northeast segment of the Keystone plate, pose an unusual problem. Suggestion has been made that the overridden blocks may represent an older thrust plate that was largely destroyed by erosion before the Keystone plate moved over the remnants (Longwell, 1926, p. 565; Glock, 1929, p. 335). This possibility of course merits consideration, but strongly against it is the fact that practically a full thickness of the Aztec sandstone underlies the Keystone plate through 12 miles of the north-south segment, up to a line marking the west limit of the tilted blocks (figs. 2, 3). It is hardly credible that erosion would remove the greater part of a thick, resistant thrust plate over a wide area and leave the underlying weak sandstone untouched, to be covered by a younger plate. Moreover it seems significant that the La Madre fault forms the abrupt boundary between the relatively simple segment of the Keystone thrust and the segment that rides over the tilted fault blocks. Between the southern exposure of the La Madre fault and Las Vegas Valley these blocks are in continuous succession, with remarkably uniform structure. The northeast dips in blocks bounded by faults that strike nearly north reflect steep plunge of the blocks northward, under the Keystone plate.

The following explanation is offered for the tilted blocks as fragments of the Keystone plate. While this plate was moving generally eastward, it was subjected to torsional strain by right-lateral movement along the Las Vegas Valley shear zone. This strain affected the entire belt now exposed between the valley and the northeast front of the high Spring Mountain Range, as indicated by strong eastward diversion of strike from the prevailing northward trend. Effect of torsion in the thrust plate is illustrated qualitatively in figure 2 by distortion of the square A B C D into the rhombus A B′ C′ D. Extension along the longer axis of the rhombus—that is, generally east-west—would produce

fractures aligned generally north-south. Blocks between fractures would be tilted eastward by continued motion of the plate and, as their own further horizontal movement would be thus impeded, they would be broken from the moving mass. Southeastward drag along the shear zone would impart a southward component to the plate, causing it to override parts of the detached blocks. Further movement of the plate would rotate the blocks to steeper dips; and strains thus induced would cause some faults between the blocks to be extended upward into the overriding plate.

Structural features in the belt of overridden blocks (figs. 2, 3) correspond well to effects expected from the assumed mechanism. Faults separating the blocks are nearly parallel, with strike nearly north as required by the assumed rhombus, which of course can have only qualitative merit. Some faults between blocks have been extended into the overlying plate, causing small and large offsets. A most conspicuous feature is the bend in strike of the Keystone thrust through 60 degrees of arc, in agreement with the general diversion in trend lines between Las Vegas Valley and the crest of the Spring Mountain Range. This diversion is attributed most reasonably to right-lateral movement in a wide zone of shearing failure. According to the analysis outlined above, relations in the belt of tilted blocks are logical incidental effects in the larger pattern of strain. Doubts may remain about the timing; were lateral movement along Las Vegas Valley and the thrusting in operation concurrently?

Pertinent features are displayed in excellent outcrops along the contact between the Keystone plate and the tilted blocks. (1) Thin-bedded units — Cambrian in the basal part of the plate, Pennsylvania in one of the overridden blocks—are buckled into recumbent folds that are overturned toward the northeast and plunge with dip of the surface. These folds give eloquent testimony to movement with a strong eastward component. (2) East of the most easterly outcrop of the Aztec sandstone in the series of tilted blocks, the base of the Keystone plate is marked by an irregular but almost continuous layer of quartz sand, mixed with particles of dark dolomite like that in basal layers of the plate. Thickness of this layer is locally as much as 8 feet. In many outcrops the sand has been forced upward into the broken plate, to form small dikes and sills. The only plausible source of the sand is the Aztec sandstone below the thrust contact in the overridden block. The layer is exposed along the base of the main plate through more than two miles; the most easterly outcrops are thin and discontinuous.

Related problems.—Therefore eastward movement of the Keystone thrust plate continued, in significant measure, after parts of the plate were broken, tilted, and overridden. If this fracturing of the plate resulted from torsional strain connected with lateral displacement along Las Vegas Valley, as argued above, the right-lateral movement was in progress during the main orogeny. The case for this verdict seems strong, when evidence from the belt of tilted blocks is joined to that given by structural relations of the La Madre and Griffith faults. But why, it may be asked, are the spectacular torsional effects confined to the Keystone thrust and not shared by other thrusts in the belt between Las Vegas Valley and the La Madre fault? The most plausible answer is that the other thrusts, as we see them, are steep, in part little more than

broken overturned folds, none of them having emerged to move over a former land surface as flat plates with large displacement. That the entire belt northeast of the La Madre fault was affected by torsional strain is suggested by numerous north-trending faults (more than shown in fig. 2) that are essentially parallel to the faults separating the overridden blocks.

Another problem concerns the abrupt termination of the tilted thrust-block fragments westward at the La Madre fault. Explanation may be twofold: (1) Maximum torsional strain may have been confined to the belt between Las Vegas Valley and the La Madre fault, which probably is a branch of the shear zone; (2) deformation by fracturing and tilting of blocks may have affected only the flat, emerged part of the thrust plate. The part of the plate in the tilted blocks had moved over a land surface, as indicated by sheared cobbles derived from peculiar rock types in the plate, deposited in channels eroded into the Aztec sandstone, and overridden by the advancing plate (Longwell, 1926, p. 570). In the part of the thrust west and south of the La Madre fault little of the emerged plate has survived erosion. A deep canyon crossing the strike shows the thrust steepening downward across upturned beds of the Mesozoic formations (fig. 3). Presumably the plate in this part of the thrust also advanced over the land surface, as far eastward as the easternmost remnants preserved in the belt of tilted blocks. Fracturing of the plate doubtless extended into the area south of the surviving remnants, where erosion has removed all traces of the thrust as well as all formations younger than early Triassic.

Another obvious question raises the possibility of branches other than the La Madre fault related to the shear zone. Conceivably the zone widens southeastward and affects a considerable belt southwest of Frenchman Mountain where all bedrock is concealed by alluvium.

MAJOR IMPLICATIONS

Some students of tectonics may find unacceptable the conclusion that right-lateral movement on a shear zone trending northwest was synchronous with development of strong folds and thrusts striking northward. A common assumption in geology has been that deformation of a geosynclinal belt results from regional compression involving the entire thickness of the crust. Adherents to this view will represent by east-west arrows the compression responsible for "crustal shortening" in southern Nevada, and will maintain that horizontal movement along a northwesterly shear zone active at the time of the compression must have been left lateral, contrary to the clear testimony along Las Vegas Valley. A general reply to this argument is that theory must be adjusted to factual evidence. How may tectonic theory be harmonized with the field evidence in this area?

The depth to which great zones of fracture extend is a matter for speculation. Some geologists suggest that lateral movement along faults such as the San Andreas may reach as deep as the Mohorovicic discontinuity, although estimates from all available data indicate that actual breakage at the time of the great earthquake in 1906 may have extended no deeper than 10 kilometers (Tocher, 1958, p. 152). Presumably the required adjustments at greater depth

are accomplished by plastic deformation. It is here assumed that the Las Vegas Valley shear zone is deep seated, and that stresses for its operation were applied below the blanket of sedimentary formations. On the other hand all available evidence suggests that folds and thrusts in this area are confined to the sedimentary section. Study of large thrusts in a wide region indicates that they were initiated (as *décollements*) along bedding above weak, thin-bedded formations, that the moving plates buckled and were sheared across, and that some emerged to move several miles over the land surface (Longwell, 1945, p. 421). No facts are known in support of a time-honored concept that such deformation involves shortening of a thick crust.

We may suppose, then, that deformation of the sedimentary formations operated under controls independent of movements in the Las Vegas Valley shear zone. Adjustments to those movements were necessary, and these adjustments resulted in the prominent bends in strike and the torsional fracturing discussed above. Under this concept the thick sedimentary blanket was essentially an inert cover into which stresses were transmitted from the active zone beneath; this would account for irregularities in the visible part of the shear zone, as the heterogeneous cover must have been distorted and the resulting fracture pattern may well show curves and branches.

But if an independent control is assumed for the strike-slip faulting, another must be invoked for the folding and thrusting in the sedimentary section. Any proposed explanation of such deformation in mountain zones must take account of the wide, comparatively thin thrust sheets which have moved horizontally many miles. Such sheets could not have been propelled simply by application of stresses to their cross sections. Whatever the ultimate cause of deformation in geosynclines, the idea of gravitational tectonics is winning a respectful hearing as a mechanism to explain some parts of the structure in mountain zones. Great blocks in the Paleozoic section east of Yellowstone Park unquestionably have moved by sliding on low-angle surfaces, perhaps with the aid of agitation from nearby volcanism (Pierce, 1957). Large detached masses in the Italian Alps appear to have a logical explanation only by gravitative movement (de Sitter, 1956, p. 277), and many students favor the view that the great nappes of the Swiss Alps were moved in part by gravity from an axial zone of uplift in the Alpine chain (e.g Jeffreys, 1935, p. 155). An impressive case is made for the importance of gravity in the structural development of the northern Apennines (Merla, 1957). Gravitative sliding is suggested as the most acceptable explanation for movement of the exceptionally wide Roberts Mountains thrust plate in north-central Nevada (Roberts et al., 1958, p. 2852).

Long a conservative in this field of thought, I am willing to consider gravitational tectonics as a significant factor in the development of mountain structure. Frictional resistance, which has appeared to be a major obstacle to extensive horizontal movements powered by gravity (Longwell, 1945, p. 427), may have been neutralized in large part by fluid pressure in the rocks (Hubbert and Rubey, 1959). Potential for large-scale gravitative adjustments may have been supplied by strong uplift in a zone parallel to the geosynclinal axis.

Van Bemmelen (1935) would have gravitative action dominant in the deformation, whereas Carey (1958, p. 332) suggests a strong lateral component of upward pressure which, acting with important gravitative movement, has caused flowage, folding, and sliding in mountain zones. Conceivably some such mechanism operated concurrently with and independently of deep-seated displacement along Las Vegas Valley.

Possible regional relations of the shear zone merit brief comment. Gianella and Callaghan (1934, p. 3) suggested that Las Vegas Valley is part of a long fault zone with right-lateral displacement, essentially parallel to the San Andreas fault and extending some 300 miles through the area near Cedar Mountain, Nevada (fig. 1), that was affected by a major earthquake in 1932. An important lineament along this course is indicated by abrupt change in trend of Basin Ranges from nearly north to northwest. The name Walker Lane was later applied to a part of this zone, not extended as far southeast as Las Vegas Valley (Locke, Billingsley, and Mayo, 1940, pl. 2). How far northwestward the zone recognized near Las Vegas may extend is a moot question. Good evidence for it is recognized as far as the north end of the Specter Range, west of Mercury, Nevada; farther northwest the projected course is covered by late Cenozoic volcanic rocks and related sedimentary beds that mantle a wide area. Conceivably the zone does extend as far as Cedar Mountain or beyond, with a course straighter than that suggested by topography. If the zone is so extensive, it may, as proposed by Gianella and Callaghan, be a close relative of the San Andreas fault. Students of the latter fault have considered the folds and thrusts in the neighboring ranges as structures confined to a shallow zone, not primarily related to forces concerned with the lateral fault movements (Hill and Diblee, 1953, p. 455).

There is a natural curiosity about the southeastward extent of the Las Vegas Valley shear zone. It is not identified beyond a cover of volcanic rocks north of the western basin of Lake Mead, but possibly it extends below these rocks and alluvial cover many miles farther. Its projected course is approximately in line with the Grand Wash Cliffs at the southwest edge of the Colorado Plateau (fig. 1); but no reported evidence of similar faulting in that vicinity is on record.

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CENOZOIC SEDIMENTATION AND CRUSTAL MOVEMENT IN WYOMING*

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ABSTRACT. One of the thickest and most complete sequences of Cenozoic nonmarine sedimentary rocks in North America is in Wyoming. During the Laramide Revolution, which, in this region, began in latest Cretaceous time and was essentially over by the close of Eocene time, some basins were downwarped 30,000 feet or more and some mountains upwarped at least 14,000 feet. Most basins were filled before the beginning of Oligocene time, and most mountain ranges were then partially buried by Oligocene, Miocene, and Pliocene strata. Some thrust faulting occurred along the southwest flank of the Gros Ventre Mountains during early Pliocene time. The highest plain of aggradation is thought to have been developed in central and northwestern Wyoming during middle Pliocene time and in eastern Wyoming in late Pliocene or early Pleistocene time. The later history of Wyoming was one of regional uplift, normal faulting, warping, and regional degradation, with minor local deposition of Pleistocene sediments.

INTRODUCTION

One of the thickest and most complete sequences of Cenozoic nonmarine sedimentary rocks in North America is in Wyoming. In preparing the summary here presented, results of the writer's more than 20 years of study of the region are freely supplemented by the published and unpublished work of

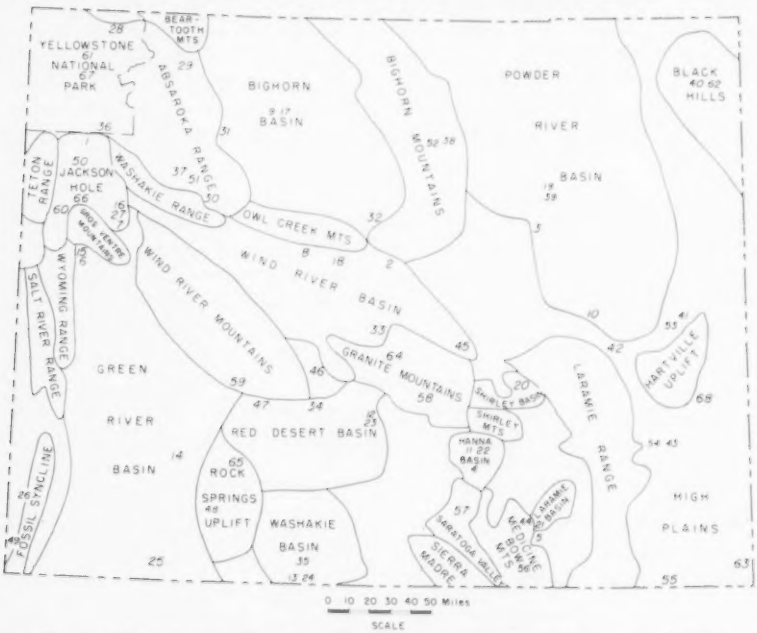

Fig. 1. Index map of Wyoming showing locations of mountain ranges and of basins containing Cenozoic deposits. Numbers indicate localities referred to in text.

* Publication authorized by the Director, U. S. Geological Survey.

many other geologists, so many in fact, that if all were properly cited, the bibliography and acknowledgments would be longer than the geologic part of the text. The reader is referred instead to excellent bibliographies for the local structural basins (fig. 1) in the Wyoming Geological Association Guidebooks, 1946-1958. Two men, however, deserve special acknowledgment for their parts in developing this story. W. H. Bradley opened the door to new and fascinating scientific vistas by his research on Cenozoic rocks in southwestern Wyoming. S. H. Knight spent a lifetime studying and inspiring others to study the Cenozoic history of the Rocky Mountains.

LATE BUT NOT LATEST CRETACEOUS

The Lewis sea (late Campanian and early Maestrichtian) was the last extensive one in which marine sediments were deposited in Wyoming. This sea covered approximately the eastern half of Wyoming, and at the time of its maximum advance, the western half of the state was an area of coastal swamps and flood plains. Before this the Mesozoic tectonic movements had been minor, in comparison with those that followed. Regional studies of thicknesses and facies of Cretaceous rocks indicate that sites of present mountains had not then risen more than minor amounts.

ONSET AND MAGNITUDE OF LARAMIDE MOVEMENT

The Laramide Revolution in Wyoming began immediately after withdrawal of the late Cretaceous sea and continued through the latest Cretaceous, the Paleocene, and into the Eocene. Movement had ceased in most areas by the close of Eocene time. During this interval, all of the major basins and all the major mountains except the Teton, Absaroka, and eastern part of the Gros Ventre Ranges were formed. The relationships of some, but only some, of the mountain uplifts to adjacent basins have remained essentially unmodified since that time. Hence, a concept of the magnitude of vertical movement during the Laramide Revolution can be obtained in the areas least complicated by post-Eocene crustal warping and faulting, by comparing present basin and mountain elevations of the Precambrian surface. Depths of the basins are based on

Basins:	Elevation on Precambrian surface with reference to present sea level datum.
Wind River Basin	minus 23,000 feet
Green River Basin	" 24,000 "
Powder River Basin	" 14,000 "
Red Desert Basin	" 23,000 "
Hanna Basin	" 31,000 "
Washakie Basin	" 21,000 "
Bighorn Basin	" 21,000 "
Mountains (amount eroded from Precambrian core since Eocene time unknown):	
Wind River Mountains	plus 14,000 feet
Bighorn Mountains	" 13,000 "
Beartooth Mountains (Wyoming part)	" 11,500 "
Medicine Bow Mountains	" 12,000 "
Owl Creek Mountains	" 9,000 "
Laramie Range	" 10,000 "

surface, subsurface, and reflection seismograph data. Those basins and mountains that were significantly affected by late Cenozoic tectonic movements are not listed.

LATEST CRETACEOUS

Broad upwarped areas in Wyoming and adjacent states furnished much of the clastic debris that was deposited during latest Cretaceous time in the downwarps in Wyoming (table 1). Two east-trending downwarps developed during the time the Lance formation (latest Cretaceous, late Maestrichtian and Danian ?) and its equivalents were being deposited. The northern one extended southeastward from the south boundary of Yellowstone National Park (fig. 1, loc. 1) into the present site of the Wind River Basin (fig. 1, loc. 2), and then northeastward (fig. 1, loc. 3) toward the Black Hills. The southern downwarp, separated from the northern one by the ancestral Granite Mountains-Wind River Mountains arch, extended eastward across the Red Desert and Hanna Basins (fig. 1, loc. 4).

The present sites of the Bighorn, Washakie, and Green River Basins sank rather uniformly and were covered by 700 to 2,000 feet of drab fine-grained sandstone, shale, and thin coal beds. In southeastern Wyoming, east of the Laramie Range (fig. 1), 1,500 feet or more of similar material was deposited. Some dinosaur-bearing conglomerate is present along the western margin of the Green River Basin (Tracy and Oriel, 1959, p. 128).

PALEOCENE

In the vicinity of the Fossil Syncline in southwestern Wyoming (Tracy and Oriel, 1959, p. 126-130), and in the Hanna Basin (Knight, 1951, p. 45-53) deposition was apparently continuous from Late Cretaceous to Paleocene time. Elsewhere, the Cretaceous-Paleocene boundary is commonly marked by a major unconformity. Before deposition of Paleocene rocks, sharp north-northwest-trending folds and thrust faults developed in and adjacent to Jackson Hole. North-trending folds were present in some places along the western margin of Wyoming. The platform area that is now the western part of the Green River Basin was tilted eastward and successively older Cretaceous rocks, from east to west, were exposed by erosion. Similarly, in the area that is now part of the Red Desert Basin, successively older Cretaceous rocks were exposed from south to north. The Wind River, Granite, and Medicine Bow Mountains continued to rise. The Beartooth, Owl Creek, western part of the Gros Ventre Mountains, the Rock Springs Uplift, Washakie Range, and Black Hills began to rise. The Bighorn Mountains either did not exist or were very low.

As broad areas, hitherto quiescent, began to rise, the first Paleocene sediments were derived from soft gray and drab Cretaceous shales and sandstones. Conglomerates, sandstones, and dully variegated claystones, were deposited adjacent to the somewhat older Wind River, Granite, and Medicine Bow Mountains. Extensive coal swamps were present in all the major basins; the greatest accumulation of coal was in the Powder River Basin.

In Jackson Hole, the Bighorn Basin, and the southwestern part of the Powder River Basin, are Paleocene conglomerates composed of quartzite roundstones derived from Precambrian rocks at an as yet unknown source probably

in or slightly west of Yellowstone National Park. The conglomerate is more than 1,000 feet thick in Jackson Hole.

A remarkable unnamed sequence of dark gray and black shale extends eastward through the Wind River Basin (Keefer and Troyer, 1956) and across the southern margin of the Powder River Basin. It is 2,000 to 4,000 feet thick and is superficially similar to marine shales of Cretaceous age. As yet, no definite marine fossils have been found in it, but shark and mammal teeth occur together in several localities. Paleocene strata were removed by erosion east of the Laramie Range and the Powder River Basin before Oligocene deposition. Therefore the possible former continuity of these shales with any part of the marine Cannonball member of the Fort Union formation of Paleocene age in South Dakota cannot be determined.

All the major mountain ranges in Wyoming were outlined by the close of Paleocene time except the north-trending part of the Teton Range, the Absaroka Range, and the eastern half of the Gros Ventre Range. All the major basins were formed, but much more deposition occurred in some than in others (table 1).

TABLE 1

Thickness in feet of uppermost Cretaceous and Cenozoic strata in Wyoming

Pleistocene	Pliocene	Miocene	Oligocene	Middle and Upper Eocene	Lower Eocene	Paleocene	Uppermost Cretaceous [Upper Maestrichtian and Danian (?)]
66[1]-400	*60*-11,500	*50*-7,000	*36*-1,500	*25*-4,000	*14*-10,000[2]	*6*-11,500	*1*- 5,000
67 -300	*61*- 2,000	*51*-2,000	*37*-3,500	*26*-1,100	*15*- 4,000	*7*- 3,000	*2*-10,000
68 - 50	*62*- 16	*52*- 150	*38*- 300	*27*-1,100	*16*- 2,600	*8*- 5,400	*3*- 3,000
	63- 170	*53*- 700	*39*- 65	*28*-1,600	*17* 9,000[2]	*9*- 6,000[2]	*4*- 6,500
	64- 1,300	*54*-1,000	*40*- 150	*29*-5,500	*18*- 6,000	*10*- 4,400	*5*- 500
	65- 200	*55*- 100	*41*- 800	*30*-2,500	*19*- 3,000	*11*- 9,200	
		56- 400	*42*-1,100	*31*-2,000	*20*- 700	*12*- 3,500[2]	
		57-2,500	*43*-1,100	*32*-1,000	*21*- 4,000	*13*- 3,300	
		58-2,800	*44*- 300	*33*- 600	*22*- 8,500		
		59- 150	*45*- 900	*34*-2,500	*23*- 7,000[2]		
			46- 500	*35*-5,000	*24*- 3,200		
			47- 700				
			48- 950[3]				
			49- 400[3]				

[1] Locality numbers (italic) before thicknesses are shown on figure 1. Numbering system for each series or subdivision of series is clockwise, beginning in western or northwestern Wyoming.

[2] Thickness estimated.

[3] Age questionable.

EOCENE

Early Eocene.—Early Eocene was the time of some of the most widespread and intense crustal deformation in Wyoming. The rise of the mountain ranges, the folding and faulting around the basin margins and within the basins, and the major subsidence of most basins were essentially completed during early Eocene time. In the central parts of most basins, gray and brown sandstone and shale and coal beds, similar to those of Paleocene age, were deposited first. Successively younger fine-grained strata are more highly variegated and the younger sandstones are more arkosic. Carbonaceous arkosic sandstones deposited during early Eocene time are host rocks for some of the largest uranium deposits in the United States. Coals are most abundant in the lower part of the sequence in the Powder River, Hanna, and Red Desert Basins.

Spectacular accumulations of granite boulders, in places more than 2,000 feet thick, are present on the basinward sides of mountain ranges that rose along thrust and high-angle reverse faults. The most extensive conglomerate of this type is along the southern margin of the Granite Mountains. One occurrence of red conglomerate less than 100 feet thick, and possibly of early Eocene age, is known east of the Laramie Range.

Cenozoic volcanic activity began in Wyoming with the development of volcanos in the southern part of the Absaroka Range. As much as 1,000 feet of tuffaceous strata, chiefly of andesitic, rhyolitic, and dacitic composition, accumulated in the northwestern part of the Wind River Basin (Keefer, 1957, p. 188-193).

At least part of the Uinta Mountains, south of the Rock Springs Uplift in Utah, began to rise during early Eocene time. Of all the major basins, only the Washakie Basin remained relatively shallow. Extensive lakes formed in southwestern Wyoming and spread around the Rock Springs Uplift from the Green River Basin through the Washakie and Red Desert Basins. Oil shales were deposited in the Luman and Tipton tongues of the Green River formation (Sears and Bradley, 1924, p. 93-107; Bradley, 1926, p. 121-131; Pipiringos, 1955, p. 100-104). Sedimentation apparently was continuous from early to middle Eocene time in the centers of most basins, although not necessarily so near their margins. Maximum thicknesses of lower Eocene rocks, including the Tipton tongue of the Green River formation are shown in table 1.

Middle and late Eocene.—Middle and late Eocene time was marked by a progressive reduction in intensity of crustal deformation. The southern margin of the Granite Mountains was upfaulted 2,000 feet or more, apparently during middle Eocene time. Erosion of the upfaulted block produced conglomerate along the northern margin of the Red Desert Basin. The Washakie and Green River Basins continued to sink after adjacent basins had been stabilized. The Uinta Mountains continued to rise during middle Eocene time along the southern margin of the Green River Basin, just south of the Wyoming state line. These mountains furnished much clastic debris that was deposited in lakes to the north.

An uplift whose nature and extent have not been determined occurred in the southern part of Jackson Hole probably at the beginning of middle Eocene time. As a result, the thick quartzite cobble conglomerate in the Paleocene

PLATE 1

Granite Mountains partially buried by Miocene and Pliocene strata. Air oblique photo looking southeast across highest peaks of Precambrian rocks in the Granite Mountains (A), rising 1,000 to 2,000 feet above Miocene (B) and Pliocene (C) strata. A surface (D) bevels Precambrian rocks above the highest remnants of Pliocene strata. Upper Cenozoic fault (E) marks southern margin of post-Miocene syncline. Ferris Mountains (F) are 30 miles away. Photo by P. T. Jenkins and L. P. House.

sequence was stripped off, moved southward by powerful streams, and redeposited as a fan 2,000 feet thick, in the northern part of the Green River Basin.

This was the time of maximum development of Eocene lakes (Bradley, 1948, p. 635-648) in southwestern, central, and northern Wyoming. In some of them were deposited oil shale, anhydrite, trona, and uranium-bearing phosphates. These lakes probably formed because drainage to the east that was established in Paleocene and early Eocene time was cut off either by rise of the eastern half of Wyoming or by subsidence of the western half, or both. No middle and upper Eocene rocks have been identified east of a line extending south from the east margin of the Bighorn Mountains to the Sierra Madre. West of that line, rocks of these ages are found in every basin and on the flanks of every Eocene mountain range. It is unlikely that all remnants of these strata would have been removed from eastern Wyoming if the magnitude of sedimentation had been comparable to that farther west.

Volcanic activity increased progressively from middle to late Eocene time. Most of the volcanic debris came from centers in the Absaroka Range and Yellowstone National Park, but smaller volcanic vents furnished local accumulations of quartz latite and alkalic rocks along the northern margin of the Granite Mountains and alkalic rocks in the Black Hills. The tuffaceous material deposited during middle Eocene time is chiefly rhyolitic, dacitic, and andesitic, whereas that of late Eocene age is petrologically a pyroxene andesite.

By the close of Eocene time, lacustrine and fluviatile tuffaceous strata had almost completely buried the Washakie and Owl Creek Mountains and the lower flanks of other ranges. The Wind River and Bighorn Basins were no longer separated. However, the Wind River, Bighorn, and Granite Mountains remained prominent.

There is no way to determine the maximum thicknesses of middle and upper Eocene strata that were originally deposited in the centers of the structural basins, because all post-Eocene strata, plus an unknown thickness of middle and upper Eocene rocks have been eroded from the central parts of these basins. Therefore, the figures given in table 1 are maximum thicknesses of incomplete sections.

OLIGOCENE

Before the beginning of Oligocene deposition major eastward drainage was reestablished in most of Wyoming. This was followed by regional degradation, and a surface of moderately high relief was cut in eastern Wyoming. In the central part of the state, pre-Oligocene erosion apparently was less vigorous.

Rocks of unequivocal Oligocene age are not known in southwestern Wyoming. However, about 950 feet of unfossiliferous white tuffaceous sandstone that may possibly be of Oligocene age is present on the southern part of the Rock Springs Uplift. This structure was raised and deeply eroded after the end of Eocene deposition and before this sandstone was laid down. Eardley (1959, p. 166-171) mapped a sequence of light-colored tuffaceous rocks that he considers to be Oligocene in the southwest corner of Wyoming. Gazin (1959, p. 137), on the other hand, believes the same rocks are late Eocene.

Similar rocks are present farther north in the Fossil Syncline (Tracy and Oriel, 1959, p. 126-130).

Enormous amounts of andesitic debris were extruded from volcanic vents along the central part of the Absaroka Range (Love, 1956a, p. 87-88) and in the northwest corner of Yellowstone National Park. Powerful streams carried away the coarse debris. Ash was transported by wind for hundreds of miles to the northeast, east, and southeast.

In contrast to the green and drab pyroxene andesitic debris characteristic of the upper Eocene, that of Oligocene age is typically gray to white and contains much more hornblende andesite. Throughout most of Wyoming, Oligocene rocks consist predominantly of white blocky claystone. Although locally derived conglomerates do occur in some places, one of the characteristic features of most Oligocene rocks is that they contain relatively small amounts of such debris, even in the places where they were deposited on surfaces of considerable relief. Apparently there was so much airborne volcanic debris settling out that the streams were overloaded and had no power to erode adjacent highlands.

By the close of Oligocene time, only the upper 1,000 to 4,000 feet of the highest peaks in the major mountain ranges (except the Teton Range, which developed its configuration much later) of Wyoming protruded above the aggradational plain. Maximum thicknesses of Oligocene rocks are given in table 1.

MIOCENE

Miocene was the time of extensive sandstone deposition in Wyoming. The most complete record of sedimentation is preserved in downfaulted areas of the High Plains, Granite Mountains, and Jackson Hole. This includes at the top, strata equivalent in age to the Barstow formation of California. Elsewhere, the record is somewhat fragmentary because of subsequent erosion of these nonresistant rocks. Volcanic vents in the Absaroka Range, northern part of Jackson Hole, and the northwest side of the Teton Range extruded large quantities of pyroxene andesite. There is 7,000 feet of tuffaceous sandstone and conglomerate in one continuous section in Jackson Hole, 2,500 feet in the Saratoga Valley, and 2,800 feet on the Granite Mountains. A rough computation of the volume of Miocene sandstone not yet eroded from the flanks of the partially buried Granite Mountains is more than 200 cubic miles. Probably a considerable amount of the sand had a volcanic origin.

Marls, algal limestones, claystones, siltstones, and sandstones were deposited in lakes in central and southeastern Wyoming. Some shales contain syngenetic uranium. Probably sand dunes were abundant but most of the sandstones show stratification more typical of fluviatile and lacustrine deposits.

Areas of mountain tops protruding above the plain of aggradation were much smaller during Miocene time than Oligocene. This plain was apparently unbroken from the Wind River Mountains southward to the state line and from the Wind River Mountains eastward to the state line, with the exception of a few of the highest peaks and ridges along the crests of the Granite Mountains and Laramie Range.

A concept of what this plain of aggradation was like during Oligocene,

Miocene, and Pliocene times can be obtained by observing a slightly modified portion of this surface 80 miles long and 20 miles wide along the partially buried crest of the Granite Mountains (pl. 1). This area was downfaulted and downwarped in such a way, probably after middle Pliocene time, that it was only moderately vulnerable to later erosion. It presents a classic illustration of the progressive burial of a major mountain range from Eocene to Pliocene time.

In order to reconstruct the original thickness and areal extent of Miocene strata in Wyoming, it is necessary to use data from northwestern Colorado (Bradley, 1936, p. 182-183; Sears, 1924, p. 279-304; Kinney and others, 1959) and northwestern South Dakota (Denson and others, 1955). Several hundred to more than 1,000 feet of Miocene rocks was probably present in southwestern Wyoming and a somewhat lesser thickness throughout northeastern Wyoming. Thicknesses in other areas are given in table 1.

PLIOCENE

At least 2,000 feet of Pliocene rhyolite flows and welded tuffs accumulated in the central part of Yellowstone National Park. Conglomerates, sandstones, claystones, marls, and diatomites are interbedded with some of the rhyolites.

During early Pliocene time, the Gros Ventre Mountains rose and the overriding block moved southwestward (Love, 1956b, p. 145-148). More than 5,500 feet of conglomerate, sandstone, claystone, limestone, and pumicite accumulated in front of these rising mountains. Middle Pliocene was a time of relative quiescence in Jackson Hole. Large lakes formed in the Hole and farther southwest along the state line. At least 5,000 feet of sediments were deposited in and adjacent to these lakes in both localities. In Jackson Hole, these sediments include more than 1,000 feet of white limestone, 500 feet of rhyolitic pumicite, 400 feet of claystone, and some diatomite. It is doubtful whether the Teton Range was more than a series of low hills at this time. However, it began to rise, probably during late Pliocene time, and at least 1,000 feet of locally derived conglomerate and rhyolitic welded tuff was deposited (Love, 1956c, p. 1911-1913).

Burial of the Granite Mountains continued in Pliocene time and at least 1,300 feet of fine-grained lacustrine and fluviatile tuffaceous sandstone, claystone, and limestone was deposited around granite knobs that marked the crest of the mountains (pl. 1, loc. A).

A remnant of fossiliferous Pliocene conglomerate and sandstone 16 feet thick has been identified in the western part of the Black Hills (B. W. Brown, unpublished M. S. thesis, Univ. of Nebraska, 1952). Similar remnants are present on the High Plains. In general, they consist of less than 100 feet of conglomerate, sandstone, tuffaceous claystone and algal limestone.

Many volcanic cones and remnants of alkalic lava flows, previously thought to be of early Tertiary age, are present in the northern part of the Rock Springs Uplift. On the basis of argon age they are now considered to be late Pliocene (S. S. Goldich, written communication to W. H. Bradley, July 15, 1959).

Several hundred feet above the highest occurrence of Pliocene strata in

the Granite Mountains are remnants of a surface that bevels Precambrian rocks (pl. 1, loc. D). This rock-cut surface probably was once continuous with the highest surface of Pliocene fill adjacent to the peaks. When the faulting and warping that involve Miocene and Pliocene rocks on the Granite Mountains are eliminated and the area is restored to its original elevation with respect to surrounding regions, this surface appears to have been continuous with one that is now between 11,000 and 12,000 feet above sea level in the Wind River Mountains (Blackwelder, 1915, p. 193-217). Similar high-level surfaces are present in the Absaroka, Beartooth, Bighorn, and Medicine Bow Mountains and are likewise believed to mark the maximum limit of burial of the mountains during Pliocene time. Development of such surfaces could only have been accomplished during a time of regional crustal stability. No such stability is indicated in the sedimentary record in northwestern Wyoming after middle Pliocene time. Uplift, normal faulting, and rapid erosion began in Jackson Hole in late Pliocene time but in eastern Wyoming it apparently was somewhat later, probably early Pleistocene. Maximum thicknesses of remnants of the Pliocene fill are given in table 1.

PLEISTOCENE

Sediments considered to be Pleistocene in Wyoming are exclusively stratified fluviatile and lacustrine deposits. During late Pliocene and early Pleistocene time, large normal faults with 10,000 to 20,000 feet of displacement developed in Jackson Hole and smaller ones formed in Yellowstone National Park, along the southern part of the Granite Mountains (pl. 1, loc. F), and the northwest margin of the High Plains. Elsewhere in Wyoming are additional normal faults that can only be dated as post-Oligocene or post-Miocene. Many of these probably are related to this period of crustal disturbance. In addition to faulting, there was both regional and local folding. Broad folds, certainly of post-Miocene and probably of Pliocene or Pleistocene age, occur in the area north of the Sierra Madre, at the south end of the Wind River Mountains, and along the Granite Mountains, north of the normal fault mentioned above. A syncline 20 miles wide, 70 miles long, and more than 3,000 feet in amplitude was formed south of the crest of the partially-buried Granite Mountains (fig. 1, loc. 58). Pliocene rocks are involved in at least some of this movement.

Regional uplift accompanied this folding and faulting, and for the first time since the beginning of the Oligocene, there was regional degradation. Pliocene river systems were superimposed across mountain ranges. Basins originally formed and filled in early Tertiary time were reexcavated, and buried mountain ranges were exhumed. Large piedmont ice masses, in places several thousand feet thick, accumulated in northwestern Wyoming and smaller glaciers developed in mountain ranges elsewhere in the state. The subsequent melting of several hundred cubic miles of ice had a profound effect on rapidity of excavation of the basin areas.

Pleistocene sediments accumulated in only a relatively few localities that were protected from regional erosion. A large lake, whose extent has not been determined, formed during some part of early Pleistocene time in Jackson Hole (fig. 1, loc. 66), probably as a result of continued downdropping of the valley

block east of the Teton Range. In and adjacent to this lake about 400 feet of relatively fine-grained sediments accumulated. Similar sediments, as much as 300 feet thick, are present in the Hayden Valley (fig. 1, loc. 67), central Yellowstone National Park (Howard, 1937, p. 92-98). Pleistocene rhyolite flows of considerable magnitude occur in the western part of Yellowstone National Park and ash deposits of similar composition are known in Jackson Hole, the Bighorn, Wind River, Red Desert, and Laramie Basins, and on the High Plains.

Pleistocene vertebrate-bearing marl, claystone, sandstone, and conglomerate, with a total thickness of about 50 feet, are preserved in a horizontal position on a hill nearly 800 feet above the adjacent Laramie River, southeast of the Hartville Uplift (fig. 1, loc. 68; L. W. McGrew, oral communication, 1959). At the time these strata were deposited, there had apparently been only a moderate amount of regional degradation. The remainder (as much as 800 feet) was accomplished during the rest of Quaternary time.

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HYDROTHERMAL ALTERATION ZONES CAUSED BY HALOGEN ACID SOLUTIONS, EAST TINTIC DISTRICT, UTAH*

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ABSTRACT. On the Helen Claim in Homansville Canyon in the East Tintic district, Utah, a small deposit of clay typical of argillic alteration, together with fluorite, diaspore, quartz, hematite, and manganese oxides, occurs at the intersection of a minor north-south fissure with a steep east-northeast trending contact of Tertiary quartz latite lava and dolomitized Cambrian Herkimer limestone. The fissure ends at a small completely argillized monzonite plug about 100 yards to the north.

Hydrothermally altered rocks are scattered through the roots of the deeply eroded Eocene volcanic field in which the East Tintic and Tintic districts lie, but in most areas of argillic alteration, abundant pyrite or alunite and barite suggest that the alteration was caused largely or in part by sulfur acids, whereas in the Helen Claim these minerals are virtually lacking.

Mineral zones parallel the lava-dolomite contact: the lava is strongly argillized in a zone about 4 feet wide, which has much montmorillonite and a small amount of kaolinite. The dolomite outward from its contact with the lava has been replaced by hematite and quartz in a zone 5-8 feet wide, which merges on the far side with a manganiferous zone 1-3 feet thick containing abundant quartz, manganese oxides and hematite, minor manganapatite, and kaolin minerals; farther from the contact, a fluorite-kaolin-diaspore zone, a diaspore-kaolin-fluorite zone, a kaolin-diaspore-fluorite zone, and a kaolin-mica zone appear successively. The edge of the dolomite is changed to a weakly bonded granular dolomite with some kaolin and mica; beyond this zone hard fresh hydrothermal dolomite extends outward for several hundred feet.

The field relations of the alteration zones, together with x-ray, mass spectrograph, and petrographic studies, suggest reaction with acid emanations carrying halides of iron, aluminum, and silicon at temperatures up to 300° C. The zoning is appropriate to precipitation responsive to an increasing pH caused by reaction with the carbonate wall rock.

INTRODUCTION

The East Tintic district is about 60 miles south of Salt Lake City, Utah. It is dominantly a terrane of middle Eocene quartz latite lavas, but many islands of Paleozoic sedimentary carbonate rocks rise above the surrounding lavas, and locally small irregular intrusive masses of quartz monzonite or monzonite cut the lavas and sedimentary rocks. The area has attracted wide interest because of the many blind ore bodies found in the Paleozoic rocks beneath the lava blanket, which has been much altered in many places. The alteration in the lava has been a useful guide to the blind ore bodies.

Some types of alteration seem closely related to ore in the underlying sedimentary rocks, whereas the other types of alteration are apparently unrelated to ore deposition as such (Lovering and others, 1949). Chloritic alteration of the lava and nearly contemporaneous dolomitization of the underlying carobnate rocks belong in this latter class—a barren-stage alteration—as does the argillic alteration of sedimentary rocks and lava.

Argillic alteration is typical of Lovering's mid-barren stage and is closely related to the intrusion of monzonite or quartz monzonite but not necessarily to deposition of ore. In most places the mid-barren stage was followed by other stages of alteration in the same locality. In the Helen Claim near Homansville Canyon, however, only the mid-barren stage is imposed on the early-barren

* Publication authorized by the Director, U. S. Geological Survey.

stage of chloritic and dolomitic alteration.[1] The locality is thus unusually favorable for isolating the effects of this mid-barren alteration stage uncomplicated by the changes induced by the late-barren and productive stages of alteration.

Argillic alteration of the mid-barren stage is commonly associated with sulfate minerals such as alunite and barite with little or no evidence of halogens, but on the Helen Claim there is little evidence of sulfur and incontestible evidence of the presence of halogens during argillic alteration.

GEOLOGY

The contact of middle Eocene quartz latite lava with hydrothermally dolomitized Middle Cambrian Herkimer limestone trends east-northeastward through a short prospect tunnel about 200 feet west of U. S. Highway 6 where it enters Homansville Canyon. The Herkimer limestone is dolomitized extensively in this vicinity, but a few hundred feet to the north it changes abruptly into normal Herkimer limestone. In the lava that nearly surrounds the island of sedimentary rock in which the Herkimer limestone is here exposed, chloritic alteration has strongly affected the basal part of the lava in the localities where hydrothermal dolomitization of the Paleozoic rocks is evident. At the prospect tunnel on the Helen Claim, about 50 feet above Highway 6, the contact of the lava and the dolomite is well exposed (pl. 1A), and both rocks have been strongly argillized. The contact is essentially an erosional contact between the lava and the underlying dolomite, but it is unusually steep and some movement has probably taken place between the two formations. In addition to minor fracturing along this contact, some fractures in the underlying dolomite trend north-northeast to a small completely argillized monzonite plug, which is well exposed at the edge of the highway about 300 feet northeast of the prospect pit. These fractures are marked by narrow discontinuous seams of white clay. The dolomite surrounding the monzonite plug contains much finely disseminated hematite and is strongly fractured and sanded.[2]

The north-northeast fissure zone is itself quite inconspicuous, but where it crosses the contact of the lava and the dolomite the rocks on both sides have

[1] Chloritic alteration is a field term that has been widely used for altered lava that has a green color suggesting the presence of chlorite.

[2] The term "sanded" as used in this paper, refers to dolomite that has a sandy "feel"; the bonds between the individual crystals have been so weakened that individual crystals loosen readily when the rock is rubbed with the finger. No silica is implied by the term sanded dolomite.

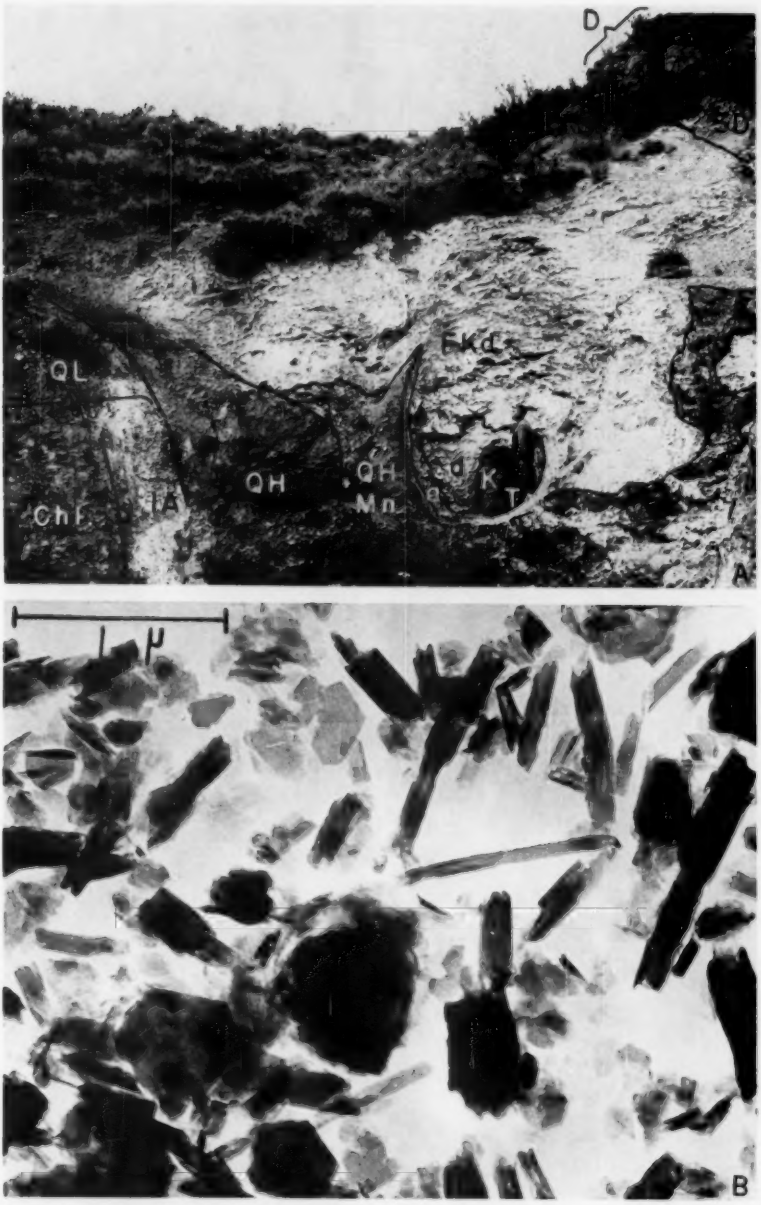
PLATE 1

Outcrop view and electron micrograph of clay at Helen Claim.

A. Contact of argillized quartz latite vitrophyre (QL) and hydrothermally altered Herkimer limestone on the Helen Claim showing alteration zones described in this paper. Chl, chloritized quartz latite; IA, intensely argillized quartz latite; QH, quartz-hematite zone (altered dolomite); QH Mn, quartz-hematite-manganese oxide zone (dolomite); FKd, fluorite-kaolin-diaspore zone; d, diaspore-kaolin-fluorite zone; K, kaolin-diaspore-fluorite zone; T, kaolin-mica transition zone; S.D., sanded dolomite; D, hydrothermal dolomite—unargillized and unsanded.

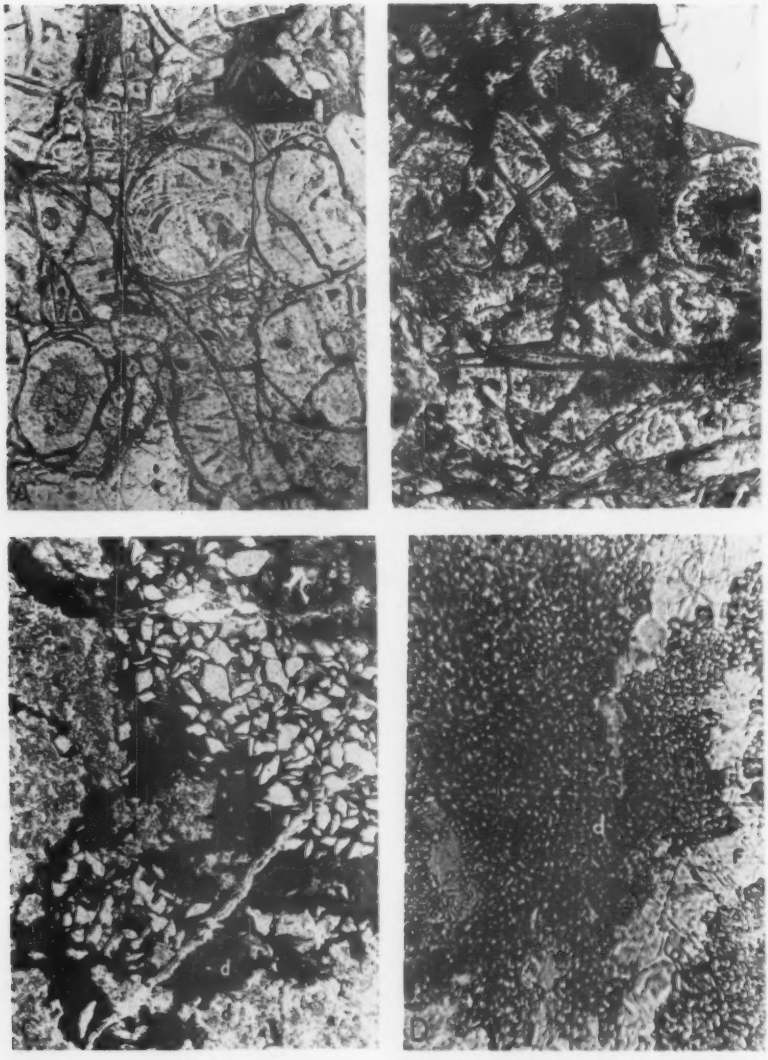
B. Electron micrograph of halloysite and kaolinite from clay seam between quartz-hematite-managnese oxide zone and the fluorite-kaolin-diaspore zone. Electron micrograph by J. C. Hathaway.

PLATE 1



Outcrop view and electron micrograph of clay at Helen Claim

PLATE 2



Photomicrographs of altered rocks from the Helen Claim

been argillized and otherwise altered along the contact for about a hundred feet.

COMPOSITION, MINERALOGY, AND PETROGRAPHY

Samples.—The general relations of the zones as seen in the field are shown in plate 1A, and the specific location of each sample referred to in the text is indicated in figure 1.

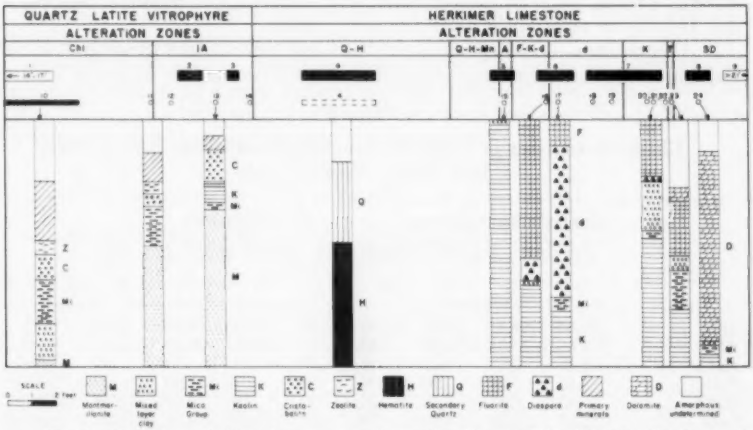


Fig. 1. Location of alteration zones and samples. Zones: Chl, chloritic; IA, intensely argillized; Q-H, quartz-hematite; Q-H-Mn, quartz-hematite-manganese oxide; A, kaolin; F-K-d, fluorite-kaolin-diaspore; d, diaspore-kaolin-fluorite; K, kaolin-diaspore-fluorite; T, kaolin-mica transition; S.D., sanded dolomite. Samples 1-9 used for chemical analysis; samples 13-20, 23-24 for optical study. Bar diagram: mineralogical composition estimated from x-ray diffraction patterns (cf. figs. 2-4).

The samples represent two series collected at different times: (a) channel samples for chemical analysis (1-9), which cover the exposure from the chloritized quartz latite vitrophyre on the outer side of the argillized volcanic rock to the hydrothermal dolomites on the outer edge of the altered carbonate rock; (b) grab samples covering the same exposure in greater detail. The optical data are based on 13 thin sections of specimens selected from the grab samples. X-ray diffraction patterns were obtained for all samples in both series.

PLATE 2

Photomicrographs of altered rocks from the Helen Claim.

A. Quartz latite vitrophyre, chlorite zone (sample 10), showing perlitic structure, zeolite alteration of margins of ovoids, and filling of cracks by secondary minerals. Plain light, × 45.

B. Intensely argillized quartz latite vitrophyre (sample 13), showing replacement of glass by montmorillonite. Crossed nicols, × 45.

C. Completely altered dolomite, fluorite-kaolin-diaspore zone (sample 16), showing early brecciated fluorite in a matrix of very finely crystalline diaspore (d) cut by a vein of late fluorite (F). The large gray areas—left side and lower right—are fine-grained diaspore (d) in a matrix of fluorite. Plain light, × 45.

D. Completely altered dolomite, diaspore-kaolin-fluorite zone (sample 18). Vein of late fluorite (F) in a matrix of diaspore (d). A triangular section of a zunyite tetrahedron (Z) can be seen within the fluorite vein, lower right. Plain light, × 190.

With one exception (sample from the quartz-hematite zone) five x-ray diffraction curves were run for each sample: (1) complete sample unoriented; (2) oriented fraction $< 5\ \mu$ or $< 2\ \mu$; the same preparation was: (3) glycolated (held in ethylene glycol vapor at 60° C overnight); (4) heated at 300° C for 30 minutes; (5) heated at 550° C for one-half hour. The quantitative estimates of the clay minerals are based mainly on the method of L. G. Schultz (in press). Finally electron micrographs were made of five samples (13, 15, 16, 21, and 24) to ascertain if halloysite were present.

Composition.—The chemical analyses and the changes in weight per unit volume of the different zones sampled across the contact are shown in Tables 1

TABLE 1

Chemical analyses in weight percent of altered rocks, Helen Claim, East Tintic district, Utah[1]

	Quartz latite vitrophyre			Altered dolomite						
	Chl	IA		QH	A	FKd	d	K	S.D.	D
	1	2	3	4	5	6		7	8	9
SiO_2	66.63	64.41	64.45	60.91	45.50	15.52		38.20	4.96	.15
Al_2O_3	14.66	15.91	16.88	6.19	32.96	52.97		39.33	3.89	.09
Fe_2O_3	2.15	2.50	2.55	22.58	2.07	.17		.54	.88	.28
FeO	.45	.00	.00	.00	.00	.07		.02	.00	.02
MgO	.98	2.47	2.46	.86	1.00	.09		1.01	17.48	21.09
CaO	2.57	1.58	1.01	.69	1.07	13.39		3.89	29.08	31.01
Na_2O	1.86	1.17	.48	.12	.17	.11		.28	.06	.02
K_2O	4.09	1.93	2.57	2.12	2.12	.42		.81	.40	.03
H_2O^-	2.29	3.47	2.55	.42	1.75	.22		1.35	.14	.03
H_2O^+	3.61	5.25	5.02	1.74	10.25	11.69		12.02	1.16	.09
TiO_2	.39	.42	.45	.29	.05	.04		.05	.04	.01
CO_2	.04	.10	.15	.04	.24	.22		.12	41.18	47.13
P_2O_5	.09	.12	.13	.34	.34	.17		.71	.04	.00(tr)
SO_3	.01	.02	.07	.08	.01	.01		.13	.00	.00
Cl	.02	.05	.02	.00	.00	.00		.05	.06	.04
F^-	.07	.29	.30	.16	.40	8.75		2.26	.20	.01
S	.00	.01	.02	.03	.02	.01		.10	.02	.01
MnO	.06	.26	.57	2.69	1.81	.01		.04	.45	.16
BaO	.17	.12	.32	.48	.10	.00		.04	.02	.01
	100.14	100.08	100.00	99.74	99.86	103.86		100.95	100.04	100.18
Less O	.03	.13	.14	.08	.18	3.68		1.01	.10	.01
TOTAL	100.11	99.95	99.86	99.66	99.68	100.18		99.94	99.94	100.17
Bulk density	2.18	1.77	1.95	2.19	2.24	2.64		2.32	2.42	2.74
Powder density	2.40	2.39	2.52	3.04	2.63	3.02		2.69	2.81	2.82
Spec										
BaO	.20	.10	.30	.30	.10	.004		.05	.02	.002
SrO	.05	.01	.01	.01	.08	.05		.50	.02	.01

[1] Analyst, N. L. Tarrant, U. S. Geological Survey, Denver, Colorado.

KEY: All samples are channel samples; see figure 1 for location of samples relative to contact of quartz latite vitrophyre with altered dolomite.
Abbreviations above the sample numbers show position of samples with respect to zones.

Zones: Chl—chloritic; IA—intensely argillized; QH—quartz-hematite; A—kaolin; FKd—fluorite-kaolin-diaspore; d—diaspore-kaolin-fluorite; K—kaolin-diaspore-fluorite; S.D.—sanded dolomite; D—dolomitized limestone.

and 2. It is obvious from the chemical analyses that sulfur is present only in very minor amounts, whereas fluorine is a major constituent of the argillized rocks. The former presence of a minor amount of pyrite in the outer zones of argillic alteration in the dolomite is shown by sparse cubic pseudomorphs of hematite and goethite after pyrite, but much of the meager brown stain in the outer zone is the result of alteration of rhomb-shaped crystals that are presumed to represent ankerite rather than pyrite. Neither barite nor alunite was observed under the microscope nor identified on x-ray patterns; it is concluded therefore that the argillic alteration in this locality is the result of magmatic emanations that were almost entirely halogen acids carrying only a trace of sulfur. Presumably their analogue would have been Fumarole No. 1 of Zies in the Valley of Ten Thousand Smokes (Lovering, 1957).

TABLE 2

Composition in weight per unit volume (mg/cc) of altered rocks, Helen Claim, East Tintic district, Utah

	Quartz latite vitrophyre			Altered dolomite					
Zone	Chl	IA		QH	A	FKd	d	K	S.D.	D
Sample	1	2	3	4	5	6	7	8	9
SiO_2	1,453	1,140	1,257	1,334	1,019	410	886	98	4
Al_2O_3	320	281	329	136	694	1,398	912	94	2
Fe_2O_3	47	44	50	495	46	4	13	21	8
FeO	10	0	0	0	0	2	0	0	1
MgO	21	44	48	19	22	2	23	423	578
CaO	56	28	20	15	24	353	90	704	850
Na_2O	41	21	9	3	4	3	6	1	1
K_2O	89	34	50	46	47	11	19	10	1
H_2O^-	50	61	50	9	39	6	31	3	1
H_2O^+	79	93	98	38	230	309	279	28	2
TiO_2	9	7	9	6	1	1	1	1	0
CO_2	1	2	3	1	5	6	3	997	1,291
P_2O_5	2	2	3	7	8	4	16	1	0
SO_3	0	0	1	2	0	0	3	0	0
Cl	0	1	0	0	0	0	1	1	1
F^-	2	5	6	4	9	231	52	5	0
S	0	0	0	1	0	0	2	1	0
MnO	1	5	11	59	23	0	1	11	4
BaO	4	2	6	11	2	0	1	0	0
TOTAL	2,180	1,770	1,950	2,190	2,240	2,640	2,320	2,420	2,740

The quantities given in table 2 (mg/cc) are calculated from the analyses and bulk densities shown in table 1; weight percent × bulk density × 10 = mg/cc. The abbreviations and the sample numbers are the same, see Key, bottom of table 1.

The most striking changes in the rocks are the decrease in bulk density with alteration, the addition of SiO_2, Fe_2O_3, MnO_2, Al_2O_3, H_2O, and F to the dolomite and the concomitant loss of CaO, MgO, and CO_2, while in the quartz latite MgO, H_2O, F, and MnO_2 were being added and CaO and Na_2O were being lost. The changes in composition reflect the fact that the chief alteration mineral in the argillized lava is montmorillonite, whereas the dolomite was

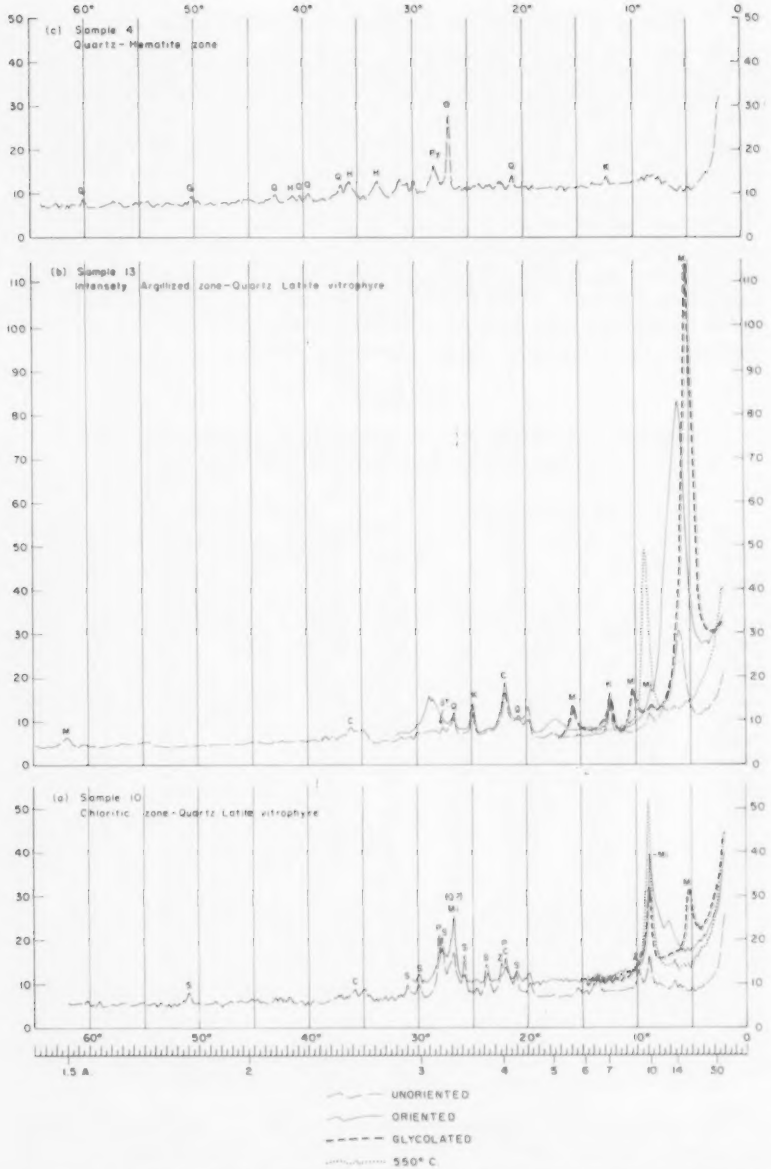


Fig. 2. X-ray diffraction curves for quartz latite vitrophyre alteration zone and the quartz-hematite zone, Cu Kα radiation. Lettered peaks: C—cristobalite; K—kaolinite; M—montmorillonite; Mi—mica group; P—plagioclase; Py—pyrolusite; Q—quartz; S—sanidine; Z—zeolite.

(a) Quartz latite vitrophyre, chloritic zone (sample 10).
(b) Quartz latite vitrophyre, intensely altered zone (sample 13).
(c) Quartz-hematite zone (sample 4).

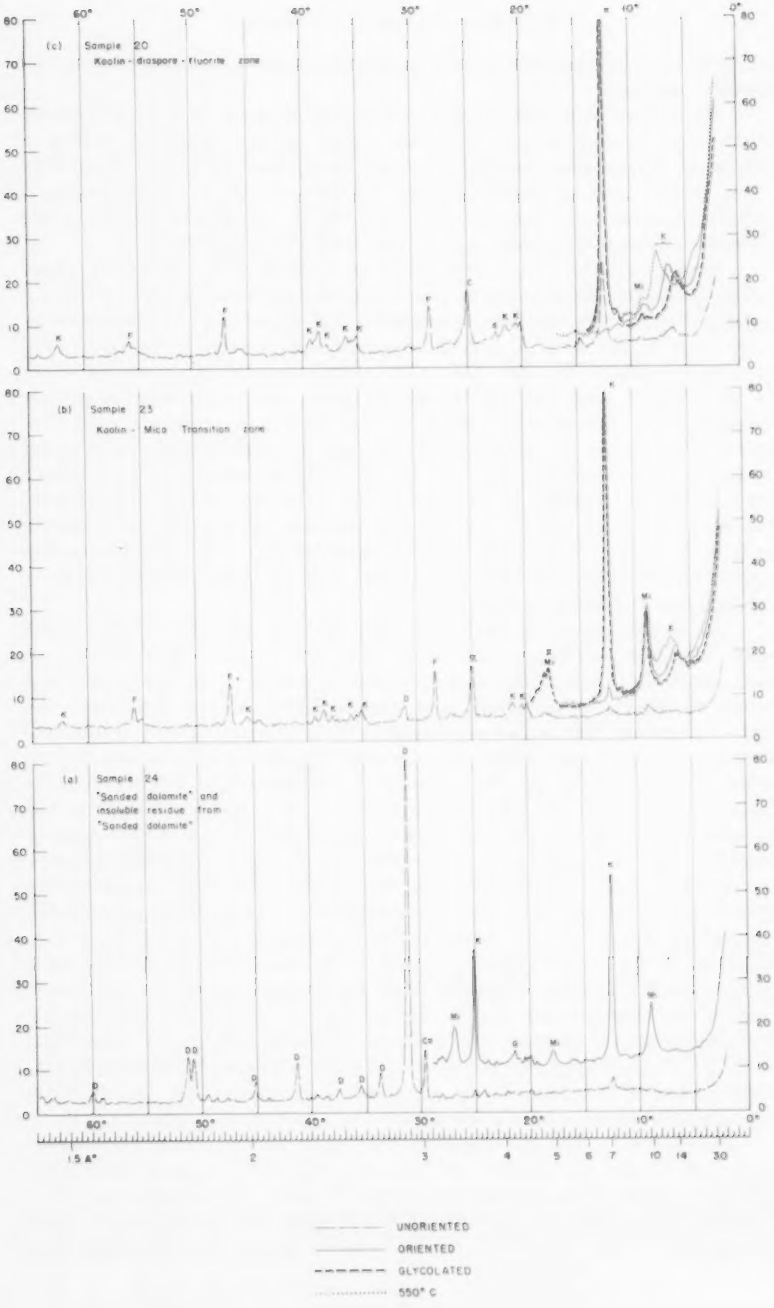
altered chiefly to hematite, quartz, manganese oxides, diaspore, fluorite, and kaolinic minerals.

Mineralogy from x-ray analysis.—The data obtained by x-ray analysis indicate the course of alteration in broad outline (see figs. 1-4). The method as employed groups some minerals that can be further differentiated optically (e.g., micas), and in general it is not sensitive enough to identify minerals present in amounts less than a few percent. On the other hand, it indicates the presence of mixed-layer clays that can not be identified optically.

Montmorillonite is the chief alteration product of the argillized quartz latite vitrophyre. In the zone of chloritic alteration, however, most of the expanding lattice clay has been interpreted as a mixed-layer montmorillonite-mica (note height of 17 Å peak as compared with increase in height of 10 Å peak after collapse, fig. 2a). The zeolite of this zone persists to within 3 feet of the contact with the quartz-hematite zone, and cristobalite is present throughout the altered latite. Only a trace of chlorite is recognized x-ray wise and is restricted to the chloritic zone. A large proportion of the mica group minerals in the chloritic zone is presumably biotite rather than a secondary 10 Å mica. The decrease in the mica group in the zone of intense argillic alteration (cf. fig. 1a and b) suggests that the mica did not alter to a mineral with 10 Å spacing. In the more intensely argillized part of the latite, beginning approximately three feet from the contact with the quartz-hematite zone, a minor amount of kaolinite appears.

The mineralogy of the altered dolomite contrasts strongly with that of the latite. Diaspore and fluorite are conspicuous in this zone and fluctuations in their abundance suggest the controlling effect of fractures. No less conspicuous is the occurrence of kaolinite as the principal clay mineral throughout the altered dolomite. The sharpness of the 7 Å peak suggests well crystallized kaolinite, but this peak masks the halloysite revealed by the electron microscope (cf. fig. 4c and pl. 1b). Mixed-layer montmorillonite-chlorite clay is present near the contact with sanded dolomite; it is high in chlorite in the transition zone but the ratio changes to approximately 1:1 in the adjacent kaolinite-diaspore-fluorite zone, where this clay increases markedly in abundance. Mixed-layer clay is absent in the samples from the diaspore-kaolin-fluorite zone, and only traces are recognized in zones nearer the quartz latite contact (see figs. 3 and 4) except in chemical sample 5 in which about two-thirds of the clay is an ordered mixed-layer mica-montmorillonite. The mica group minerals fluctuate in abundance in the altered dolomite but are prominent in the clay mineral fraction of the sanded dolomite, the kaolin-mica transition zone, and in sample 19 from the diaspore-kaolin-fluorite zone (figs. 3 and 4). Of particular interest is a 6.15 Å peak in sample 21 from the kaolin-diaspore-fluorite zone, tentatively identified as boehmite (fig. 3c). Goethite was identified in the sanded dolomite and the transition zone adjacent to it.

Petrography.—The mineralogical changes in the lava from the unargillized rock toward the contact with the quartz-hematite zone of the altered dolomite, as seen under the microscope, are instructive: The slightly chloritized lava that is the host rock of the argillic alteration is a perlitic quartz latite vitrophyre (pl. 2A). It is chiefly a light brown to almost colorless perlitic glass


(c) Sample 20
Kaolin - diaspore - fluorite zone
(b) Sample 23
Kaolin - Mica Transition zone
(a) Sample 24
"Sanded dolomite" and insoluble residue from "Sanded dolomite"
60°
50°
40°
30°
20°
10°
0°
1.5 A°
2
3
4
5
6
7
10
14
30
UNORIENTED
ORIENTED
GLYCOLATED
550° C

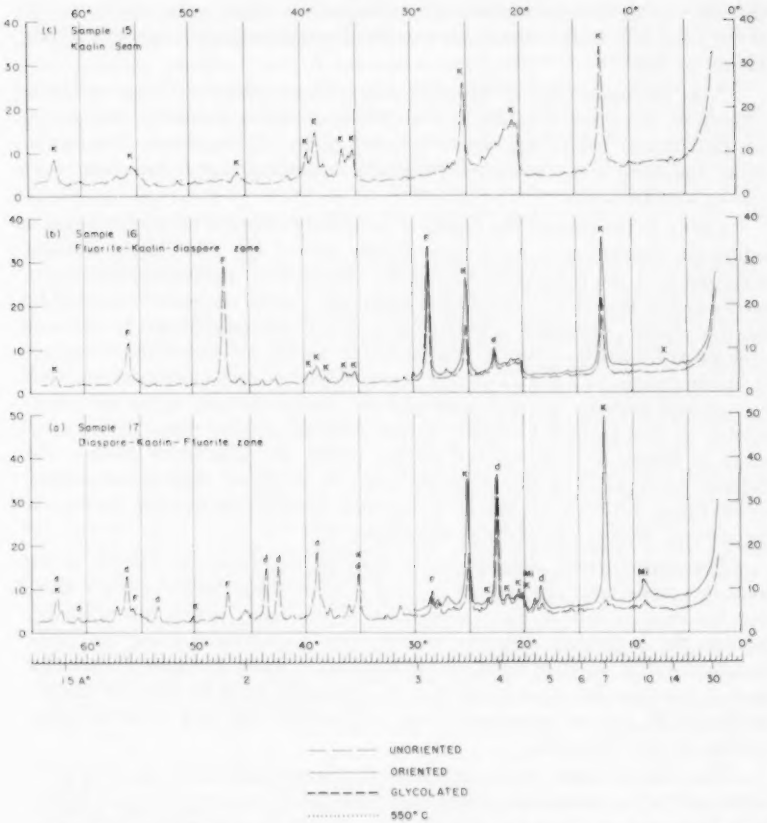


Fig. 4. X-ray diffraction curves for inner part of altered dolomite zone. Cu Kα radiation. Lettered peaks: d—diaspore; F—fluorite; K—kaolinite; Mi—mica group; X—x-layer clay.

(a) Diaspore-kaolin-fluorite zone (sample 17).
(b) Fluorite-diaspore-kaolin zone (sample 16).
(c) Kaolin seam (sample 15).

which contains many phenocrysts of fresh andesine, sanidine, quartz, magnetite, and biotite, with minor apatite and brown hornblende. The glass is somewhat altered along the perlitic cracks and at the edges of ovoids to chlorite, zeolite, and cristobalite—minerals that are characteristic of the early-barren stage of

Fig. 3. X-ray diffraction curves for sanded dolomite and adjacent alteration zones. Cu Kα radiation. Lettered peaks: Ca—calcite; D—dolomite; d—diaspore; F—fluorite; G—goethite; K—kaolinite; Mi—mica group; X—x-layer clay.

(a) Sanded dolomite and its insoluble residue (sample 24).
(b) Mica-kaolin transition zone (sample 23).
(c) Kaolin-diaspore-fluorite zone (sample 20).

alteration in the district. Some montmorillonite is present in the perlitic cracks of the glass, but neither the ovoids nor the phenocrysts present are appreciably altered to clay.

Near the outer edge of the zone of argillic alteration of the quartz latite vitrophyre, the glass, chlorite, and mixed-layer clay are entirely changed to montmorillonite before the zeolite is more than half argillized. The biotite shows bleaching and a lower birefringence, suggesting that it has changed in part to a hydromica.[3]

Closer to the contact the zeolite is completely changed to montmorillonite before the plagioclase is more than slightly altered (pl. 2B). In the strongly argillized lava the plagioclase first shows alteration to montmorillonite along cracks and in the interior of zoned crystals, and then is completely replaced by the clay. Some cristobalite is associated with the montmorillonite. In the most strongly argillized lava only quartz, sanidine, magnetite, and a minor amount of biotite remain in a matrix of clay that is chiefly montmorillonite. Some very fine-grained kaolinite is intergrown with the montmorillonite, and a few minor veinlets of fine-grained kaolinite cut the montmorillonite. Some of the magnetite is altered to hematite but neither pyrite nor goethite is present. An electron micrograph of clay from the intensely argillized zone shows no halloysite tubes. Endellite, if present, would show as halloysite because the heat of the electron microscope partially dehydrates it.

The hydrothermal dolomite more than 50 feet from the contact of the lavas is a moderately coarse even-grained dolomite, and the microscope shows that the rock is full of dusty inclusions of low index which are dispersed through the individual dolomite crystals and concentrated along the grain boundaries. The negative relief of this material in those dolomite crystals that show ε' less than balsam suggests that the inclusions have an index of slightly less than 1.50, and we have classed them as allophane. The rock is cut by minor veinlets of clear dolomite.

The fresh dolomite grades into sanded dolomite, which is somewhat brecciated and full of microvugs. Some vugs have calcite overgrowths on euhedral dolomite crystals and in some vugs there are fluorite crystals several millimeters in diameter, but the relation of fluorite to calcite was not seen. In vugs lacking fluorite, however, the succession is apparently first calcite, then kaolinite and allophane, and then hydromica. Laths of an isotropic mineral, the index of which is approximately 1.60 are moderately abundant along the cleavage of the dolomite; this mineral is tentatively identified as a chlorite. Iron and manganese oxides are common along micro-fractures and as a replacement of

[3] Hydromica in this paper refers only to the mica of lower birefringence and index than muscovite—but $\alpha > 1.54$—as identified in thin section; the micas and related minerals identified by x-ray that have a basal spacing of about 10 Å and are generally reported as "illite" (which may include biotite, muscovite, hydromica, and other micas) will be termed mica group. In this paper the term mica group implies x-ray identification based on the 10 Å peak. Halloysite ($Al_2SiO_5 \cdot 2H_2O$) as used here is an isotropic clay ($n = 1.55$ to 1.556) that is formed from endellite ($Al_2SiO_5 \cdot 4H_2O$) by the loss of $2H_2O$. It is intimately intergrown with well crystallized, very fine-grained kaolinite, and poorly crystallized kaolinite, but the mixture appears isotropic under the petrographic microscope. Halloysite, as identified petrographically in the samples from the Helen Claim, probably everywhere consists of such an isotropic mixture of both kaolinite and halloysite in varying proportions.

individual carbonate grains in local areas. Hematite replaces nests of diamond-shaped crystals and is probably pseudomorphic after ankerite. The hematite is surrounded by goethite stains. A manganese oxide mineral is present in minute stubby prisms 2 to 4 microns in length, which have the crystal shape of manganite. The chemical analyses (table 1) indicate that the hematite and manganese oxide mineral seen in thin section comprise less than 1 percent of the sample—amounts too small to be detected in the x-ray analyses. No halloysite was identified by x-ray in the insoluble residue from the sanded dolomite but electron micrographs show the typical tubular crystals of halloysite ($Al_2SiO_5 \cdot 2H_2O$).

A thin gradational contact separates the sanded dolomite from strongly argillized rock that is dominantly clay. A thin section from this kaolin-mica transition zone, which contains only a very small amount of carbonate, shows mostly halloysite and fine-grained fluorite. This halloysite, index slightly higher than balsam, was previously identified in thin section as endellite by Lovering during his studies of alteration in the district, but without supporting evidence from x-ray or the electron microscope. The halloysite is crowded with minute inclusions of allophane and isotropic chlorite (?) and is the matrix for the fine-grained fluorite. Scattered through this halloysite-fluorite groundmass are a few minute flakes of montmorillonite, and many flakes of what has been termed "beidellite" in previous alteration studies in the district ($\alpha < 1.54 < \gamma$, $\gamma - \alpha \equiv .025$, Lovering and others, 1949), but which we believe to be the mixed-layer chlorite-montmorillonite mineral identified by x-ray. This halloysite-fluorite groundmass is cut by veinlets of kaolinite (grain size about 10 to 20μ). A vein of relatively coarse fluorite cuts both matrix material and kaolinite veinlets and is itself cut by a vein of kaolinite that contains the mixed-layer clay in well oriented crystals which show a crisscross pattern. Adjacent to this vein the halloysite-fluorite groundmass is replaced locally by kaolinite and hydromica; all are cut by a veinlet of clear halloysite and kaolinite. Manganese oxide follows the edge of the kaolinite-mixed-layer clay veinlet and is apparently contemporaneous with it. Some hematite pseudomorphs after cubic pyrite are sparsely distributed in this rock and are associated with late goethite stains. This is the only material studied in which evidence of the former presence of pyrite was seen.

The mineralogy changes from the transition zone to the contact of the argillized lava, but there is everywhere evidence of many stages of repeated waves of alteration. Fluorite, which is one of the earliest minerals, appears in many generations as do the aluminous minerals (pl. 2C). Some veinlets of clear fluorite, which cut diaspore in the diaspore-kaolin-fluorite zone (No. 18, fig. 1), contain zunyite, Al_3SiO_5 (OH, F, Cl)$_3$, in the center of the veinlets (pl. 2D). From the transition zone toward the lava contact, the earliest mineral is usually a much corroded fluorite that is partly replaced by diaspore, but in the fluorite-diaspore-kaolinite zone the early fluorite is partly replaced by fluellite ($AlF_3 \cdot H_2O$). Commonly a brecciated early fluorite lies in a matrix of halloysite (verified by electron micrographs of clays from zones SD, K, d, FKd, and A, fig. 1), a mixed-layer clay (?), and well crystallized kaolinite in which minor chlorite ($<2\mu$) and some later well oriented laths ($>2\mu$) of either

sericite or hydromica are present. The matrix mineral seen under the microscope is tentatively identified as the mixed-layer clay found in the x-ray spectrographs but is difficult to distinguish from hydromica. It occurs as a fine-grained mosaic of irregular flakes 1μ or less in diameter, the mean index is above that of halloysite and the lowest index is between the indices of balsam and halloysite. Hydromica flakes are locally abundant in this matrix and apparently replace it.

The general age relations described above are displayed in all the zones between the sanded dolomite and the quartz-hematite-manganese oxide zone. Although fluorite, diaspore, and kaolinite may be found in almost any specimen, the relative abundance of the minerals changes toward the contact with the argillized lava (see fig. 1).

A late kaolin seam about 6 inches wide (sample 15, fig. 1), which consists largely of halloysite and kaolinite, separates the fluorite-rich zone from the manganese-rich edge of the quartz-hematite-manganese oxide zone. A mixed-layer clay of moderately high birefringence is locally abundant in the kaolinite-halloysite matrix and is oriented in a crisscross pattern; the matrix is cut by a few cross-fiber veinlets of coarser (about 50μ) mixed-layer clay intergrown with sericite. The low relief of the isotropic halloysite against balsam suggested a study with the electron microscope, which established the presence of the tubular crystals of halloysite (pl. 1B).

The edge of the quartz-hematite-manganese oxide zone next to the halloysite-kaolinite seam contains some well formed manganapatite crystals which include minute pyrolusite "roses" and discrete inclusions of manganese oxide. Much of the pyrolusite, however, is later than the manganapatite and silica. Some coarse crystals of the dark silvery gray oxide were handpicked, x-rayed, and analyzed (table 3); they proved to be an unusually pure pyrolusite. The

TABLE 3

Semiquantitative spectrographic analysis (in percent) of crystalline pyrolusite from quartz-hematite-manganese oxide zone, Helen Claim[1]

Si	3.	P	0.0	Cu	0.015
Al	0.15	Mn	major	Pb	< 0.005
Fe	1.5	Ba	0.15	Zn	< 0.02
Be	0.00015	Sr	0.015	Y	0.007
Mg	0.015	Co	0.007	Yb	0.0007
Ca	0.3	Ni	0.003	W	< 0.01
Na	0.7	Cr	0.00015	As	< 0.1
K	0.7	Mo	0.03	Sb	< 0.01
Ti	0.015	V	0.015		

All other of the 69 elements looked for in the above semi-quantitative spectrographic analyses were absent or below the limits of detection.

[1] Spectrographer, A. T. Myers, U. S. Geological Survey, Denver, Colorado.

pyrolusite diminishes greatly toward the quartz latite vitrophyre contact and concomitantly the quartz-hematite-manganese oxide zone grades into the quartz-hematite zone. The jasperoid is made up chiefly of normal quartz but

chalcedony has replaced cubic crystals (fluorite ?) and is vuggy. In the quartz-hematite-manganese oxide zone some of the vugs are lined with pyrolusite and halloysite. Most of the hematite appears to be contemporaneous with the earlier quartz and chalcedony that makes up the matrix of the jasperoid, but veinlets of clear quartz and pyrolusite cut the early hematite-quartz intergrowth.

TEMPERATURE OF FORMATION

Quartz and hematite from the center of the quartz-hematite zone were studied in the laboratories of the California Institute of Technology by Engel and his associates. Engel reports (letter of May 14, 1956, to H. T. Morris) that the ratio of oxygen isotopes in the two minerals suggests "temperatures well above 200° C, probably above 300° C." At temperatures near 300° C the diaspore field as shown by Kennedy (1959), would require a pressure of only 100 to 200 bars of water vapor; higher temperatures would result in corundum unless the vapor pressure of water were increased greatly. Lower temperatures would tend to produce metastable boehmite—especially if less than 200° C. Boehmite however was found only in a sample from the kaolin-diaspore-fluorite zone two feet from the sanded dolomite.

Roy and Osborne (1952) and Bates (1952) agree that endellite is unstable above a temperature of 170° C even at steam pressures of 1000 bars. Endellite was presumably an early "fringe zone" alteration mineral in the cooler part of the reaction zone and was replaced by halloysite, kaolinite, or diaspore at higher temperatures. The 6-inch halloysite-kaolinite seam between the fluorite-diaspore-kaolinite zone and the quartz-hematite-manganese oxide zones is thermally incompatible with its walls and suggests a later origin from cooler solutions.

CONCLUSIONS

The mineralogy and the paragenesis of the argillized lava and dolomite indicate that the alteration was accomplished by halogen-acid solutions low in sulfur and rich in fluorine. The dolomite was largely removed but a substantial amount of calcium was fixed as fluorite. The hot acid fluorine-bearing solutions carried a substantial amount of silica, iron, and aluminum which were precipitated zonally in various minerals as the solutions reacted with the dolomite.

The controlling factor in the formation of kaolinite rather than montmorillonite in the dolomite seems to have been the silica-alumina ratio of the solutions, as both magnesium and calcium must have been dissolved in large quantities from the dolomite. In contrast, the altered lava adjacent to the dolomite contains abundant montmorillonite but little kaolinite. Only a minor amount of magnesium and other bases are present in the montmorillonite zone, but silica makes up two-thirds of the rock by weight.

It seems probable that the solutions at the intersectnon of the northeasterly trending fissure and the contact were markedly acid and that as they moved out toward the sanded dolomite they became less acid and perhaps neutral or even alkaline in the outermost zone of alteration. The comparatively sharp contact between the quartz-hematite-manganese oxide zone and the

fluorite-diaspore-kaolin zone is believed to reflect an increase in the pH of the solutions, and the hydrolysis of the iron, manganese, and aluminum successively with decreasing acidity of the hydrothermal solutions. The sharp increase in hydromica and in the proportion of chlorite in the mixed-layer clays immediately adjacent to the sanded dolomite is presumably due to both the change in pH and the increased concentration of magnesia in the solution at this place.

The many generations of the various hydrothermal minerals in the altered dolomite suggest—but do not prove—a fairly long period of hydrothermal attack, and the hydrothermal solutions may have changed in composition during the time represented by the alteration. If so, the quartz-hematite-manganese oxide zone may have formed from solutions which differed in composition from those that caused the argillic alteration of the dolomite and introduced fluorite. There is no concrete evidence, as yet, that the quartz-hematite alterations belongs to a different episode from that of the argillic alteration. No veinlets of quartz and hematite were found cutting the fluorite-diaspore-kaolinite zones at a distance, as would be expected if the quartz-hematite alteration were late. On the other hand, vugs in the quartz-hematite-manganese oxide zone contained halloysite and kaolinite which suggests that late stage solutions belonging to the argillic alteration stage were still active in this area after the quartz-hematite alteration had ceased. Moreover, the halloysite clay seam between the quartz-hematite and fluorite-diaspore-kaolinite zones clearly establishes the lateness of some of the argillizing solutions. It is believed that the evidence now available favors the interpretation that the quartz and hematite belong to an early high temperature stage of an alteration episode that was characterized by the precipitation of endellite in outer zones throughout its duration, and that endellite was therefore deposited at a late stage in zones that had formerly been above its decomposition temperature. The change from endellite to halloysite is in part recent—a near surface dehydration caused by the semiarid climate— but endellite in the outward moving fringe zone of alteration must have changed to halloysite where overtaken by the expanding zones of more intense alteration and high temperatures. Endellite probably is not present now at the surface as all the x-ray curves lack its distinctive pattern.

The composition of the hydrothermal solutions was no doubt affected by reaction with wall rocks at depth, but the halogen anions must have been of magmatic origin. Much of the silica, alumina, iron, and manganese introduced into the dolomite, however, could have been displaced from monzonitic intrusions at depth during argillic alteration, which nearly everywhere caused a marked loss in these constituents when measured in weight per unit volume (Lovering and others, p. 46, 50, 1949).

The manganese oxides in the quartz-hematite-manganese oxide zone are of especial interest because of their probable hypogene origin. The inclusions of black oxide in the interior of the manganapatite crystals indicate that the oxide is in part contemporaneous with that mineral, but the pyrolusite associated with the late veins of halloysite shows that its period of deposition outlasted that of the apatite and that it was in part deposited below a temperature

of 170°C, temperatures at which endellite may form. The intergrowth of the manganapatite with the quartz and hematite of the jasperoid, and the lack of apatite in late veinlets, suggests that the three minerals are nearly contemporaneous and therefore hypogene. The development of the sparse goethite pseudomorphs after pyrite in the thin transition zone next to the sanded dolomite could not have generated much acid, and we believe that only a small part of the manganese oxide present is related to hydrolysis of supergene acid sulfate solutions. On the other hand, it is possible that some—or all—of the meager manganese in this narrow kaolin-mica transition zone is derived from supergene bicarbonate solutions that moved through the porous sanded dolomite. The ratios of O^{18} to O^{16} in the manganese oxides and associated contemporaneous minerals may answer questions as to their origin in different parts of the argillized dolomite, but such work has not been done. The field relations and mineralogy indicate that a substantial part of the total pyrolusite is hypogene.

ACKNOWLEDGMENTS

The x-ray analyses were made under the supervision of L. G. Schultz. Apparent discrepancies between x-ray and petrographic data on the kaolin minerals were resolved by electron micrographs made through the courtesy of J. C. Hathaway. We are deeply indebted to both of them for many fruitful discussions of questions raised by the x-ray analyses.

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CYCLES IN CARBONATE ROCKS

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ABSTRACT. Sedimentary cycles are readily recognized in rock sequences involving both marine and continental strata and to a lesser extent in strata that are entirely marine but include both carbonate and detrital types. In rock sequences composed entirely of carbonate rock, cycles commonly can be detected by trends in the size and abundance of terrigenous components. In carbonate formations that are largely devoid of detrital minerals, however, cycles must be established on the basis of differences in size of clastic carbonate grains which may indicate differences in turbulence but are not necessarily the results of transportation. Trends in distribution obtained from a detailed analysis of cycles, however, may indicate whether the position of various size grades has been determined by transportation or by environment.

NATURE OF SEDIMENTARY CYCLES

A cycle as applied to the sedimentary record in geologic history is described as a recurrence, repetition, or return to a starting point, repeated at more or less irregular intervals (Weller, 1930, p. 99). The significance of sedimentary cycles of various types and magnitudes was discussed many years ago in a carefully considered analysis by Barrell (1917, p. 745-904) and cycles have since been described and discussed in many papers from many different approaches.

During the 1930's studies of cyclic sedimentation dealt largely with late Paleozoic sequences commonly referred to as cyclothems (Weller, 1931; Moore, 1931; Wanless and Shepard, 1936). These sequences are composed of comparatively thin but sharply differentiated rock units, some of marine and others of continental origin, considered to be the result of changes in relative sea level. The ultimate cause of such changes is still not generally agreed upon, but apparently it was responsible not only for many repetitions of each depositional environment, but also for very extensive developments of each. Probable depth limits of various units within the marine deposits of a Permian cyclothem have been estimated by Elias (1937) through comparison of fossil assemblages with their nearest living counterparts of known depth range.

More recently, and especially during the past decade, cyclic sedimentation has been recognized in rocks that are entirely of marine origin. The cycles are not nearly so obvious as in formations where beds of continental origin, including coal, alternate with arenaceous, argillaceous, and carbonate beds containing marine fossils; nevertheless, cyclic changes in environment are demonstrable. In southern Arizona, for instance, a marine sequence of limestone, mudstone, and some sandstone comprising the Naco group of Pennsylvanian and Permian age exhibits alternations of rock type representing cyclic development (Wanless and Patterson, 1952). These rocks are classified into textural types that can be arranged in a sequence of apparently increasing depth zones, and from them at least 70 cycles are recognized by Ziebell.[1]

Though less apparent, cyclic sedimentation may also be recognized in rocks entirely within carbonate sequences. The cycles are recognizable through comparison of included faunas and detrital grains. Subtle changes as repre-

* Publication authorized by the Director, United States Geological Survey.

[1] Ziebell, W. G., 1955, Interpretation of Pennsylvanian sedimentation from textural studies of strata at Superior, Arizona: Illinois Univ., M. S. thesis, 124 p.

sented by differences in grain diameter (clasticity), frequency (number of quartz particles present in a given surface of thin-section), or in the abundance of particular organic components, give evidence of cyclic development. This field has been pioneered by Carozzi (1950, 1954a, 1954b) who has developed techniques for obtaining statistical measurements and for graphic representation of trends; he has demonstrated the applicability of his methods even to partly dolomitized rocks.

Where grains of quartz and other terrigeneous minerals are very scarce or locally absent in sequences of limestone, cycles may be recognized by changes in carbonate grain size. Such changes, however, unlike those among land-derived detrital particles, are not necessarily related to amount of transportation. Grains of limestone, even where clearly of clastic origin, may have been derived from nearby beds as intraformational deposits or from organisms that lived in the vicinity. Belts of sediment in the quiet waters in deeper parts of Kapingamarangi lagoon (McKee, Chronic, and Leopold, 1959) furnish an illustration of particle size that has been determined by environmental control of organic distribution rather than by current or wave transportation. In this lagoon each sedimentary belt or zone is recognized by distinctive clastic (mostly bioclastic) particles of limited size range, but the grain-size characteristics of each belt bears no relation to its bathymetric position.

REDWALL LIMESTONE AS A CYCLICAL DEPOSIT

The Mississippian Redwall limestone of the Grand Canyon, Arizona, consists largely of massive carbonate beds that superficially seem to be uniform in texture throughout and, as shown by insoluble residue analyses, are remarkably free of noncarbonate detritus. Detailed sampling, however, has demonstrated that this formation is not uniform in texture but contains, within individual massive units, sedimentary cycles represented by differences in grain size. Grains range in size from cryptocrystalline to very coarse.

In the upper two members of the Redwall limestone very little recrystalization or dolomitization has occurred, so samples for textural analyses were taken at five-foot intervals in each of a series of eight sections. These samples, with the exception of aphanitic varieties in which grains are too small to examine petrographically, are composed of clastic particles of three principle types—bioclasts, intraclasts (defined as intraformational clastic fragments), and oölites. The clasts are largely unmixed in some rock units, but are mixed in most. Only locally has the texture been obscured through dolomitization. Grain size differences as represented in figure 1 reveal that five cycles are represented in the upper part of the Redwall; they have been recognized throughout the length of the Grand Canyon or for more than one hundred miles in an east-west direction.

Although grain size in clastic carbonate particles may bear no relation to depth or distance from shore, two features in the Redwall limestone suggest that in this formation the cycles may result from systematic changes in depth as reflected by difference in wave base. First, within each cycle, rocks of intermediate grain size complete a progression from one extreme in size to the other; the rocks do not follow a random arrangement in size grades. Second,

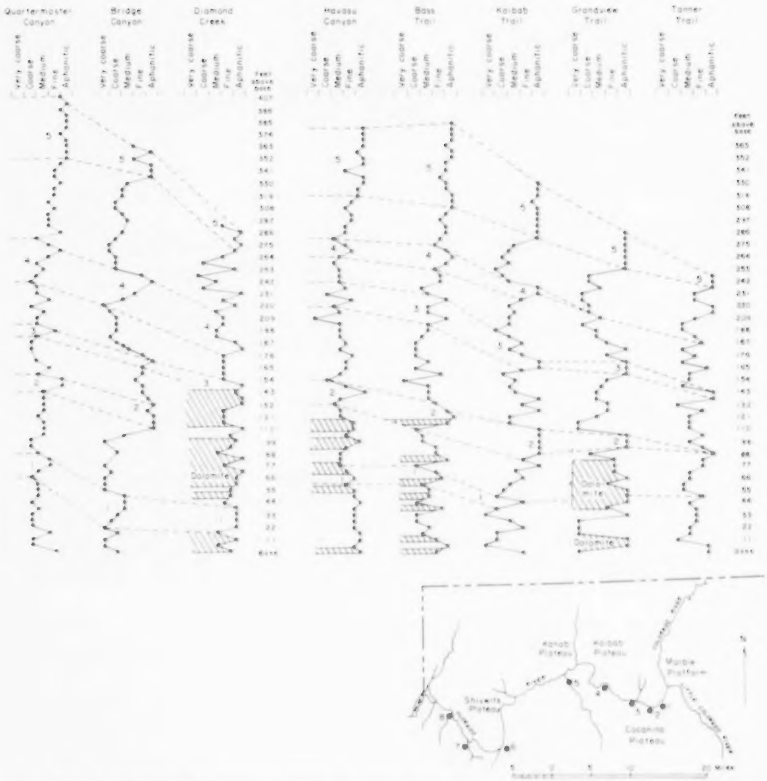

Fig. 1. · Curves illustrating cycles in grain size of carbonate rocks forming upper part of Redwall limestone at eight localities in Grand Canyon, Arizona. Numbers 1 through 5 alongside each curve represent stages characterized largely by aphanitic or very fine grained texture. Numbers at top refer to positions on locality map. Dolomite which replaced limestone and thus locally obscures the original texture is shown by cross hatching.

sections in the eastern part of the Grand Canyon contain within each cycle relatively more fine-grained clasts and do not attain the extremes in coarseness found in sections farther west (fig. 1). Thus, these cycles suggest a history of alternate changes in depth above and below wave base, probably the reflection of transgressions and regressions of a shallow sea over a broad even floor.

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RELATION OF RESERVES OF THE ELEMENTS TO THEIR CRUSTAL ABUNDANCE

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ABSTRACT. The tonnage of minable reserves in short tons (R) for many elements in the United States is equal to crustal abundance in percent (A) times 10^9 to 10^{10}, and all 26 elements for which data are available have ratios in the range $R = A \times 10^8$ to 10^{11}. The linear relation that appears to prevail between reserves and abundance for all except the most abundant elements is useful in forecasting reserves in large segments of the earth's crust or over the world at large. For purposes of estimating world reserves of unexplored elements a figure of $A \times 10^{10}$ to 10^{11} probably will give the right order of magnitude.

An interesting and surprising relation exists between the minable reserves of various elements in the United States and the abundance of the same elements in the earth's crust. The tonnage of minable reserves in short tons (R) for nearly half the 26 elements on which data are available is equal to crustal abundance in percent (A) times 10^9 to 10^{10}; about one-fourth of the others have ratios within the range $R = A \times 10^8$ to 10^{11}, and all have ratios within the range $R = A \times 10^6$ to 10^{11} (fig. 1). The relation of reserves to abundance appears to be nearly linear, except perhaps in the range of abundances greater than 1 percent, where tonnage appears to increase at a rate that is less than directly proportional to increase in abundance.

Two aspects of this relation are unexpected—its linearity over most of the abundance range, and the high correlation between reserves and abundance for all the elements on which data are available. The mere facts that the known reserves of abundant elements such as iron are large, and the reserves of rare elements such as gold are small are enough to suggest that crustal abundance is one of the factors that determines the tonnage of minable reserves of a given element; but as the concentration factors (that is, the ratio between grade of the ore and abundance of the element in the crust) of minable ores range from about 3 for the more abundant elements to nearly 10,000 for the rare ones, we might have expected that the relation of abundance to tonnage would have been an exponential one. And even if we assumed that some mathematical relation does exist in nature between abundance and tonnage, we might have expected that this relation would have been concealed for many elements by other factors. For example, if the area studied were not a representative sample of the earth's crust, the relation between reserves and abundance would be cloudy, if evident at all. Differences in the geochemical behavior among the elements (e.g. the tendency of some to be dispersed in common minerals rather than concentrated in ores), differences in the amenability of the elements to artificial concentration, and differences in the extent to which they have been sought and explored might conceal a close relation between reserves and abundance, even if one existed. What, then, is the explanation of the close relation shown in figure 1 between reserves and abundance?

Examination of the factors that might conceal a relation between reserves and abundance indicates that they do not apply to an appreciable extent to most of the elements for which reserve data are available in the United States.

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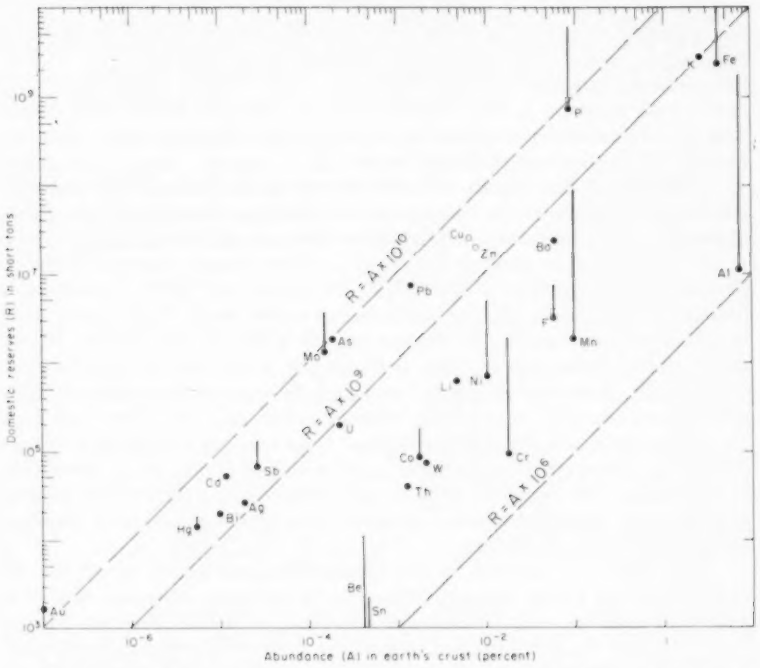

Fig. 1. Domestic reserves of elements compared to their abundance in the earth's crust. Tonnage of ore minable now is shown by a dot; tonnage of lower-grade ores whose exploitation depends upon future technological advances or higher prices is shown by a bar. The abundance of tungsten is from Green (1953); all other abundance data are those of Vinogradov (1956) for a crust consisting essentially of two parts of acidic rocks and one part of basic rocks. The estimates of reserves of uranium are from J. C. Johnson (1959), and those on potash were furnished by C. L. Jones of the U. S. Geological Survey. All other reserve estimates were prepared by members of the U. S. Geological Survey and U. S. Bureau of Mines and published in Mineral Resources of the United States (1948), the President's Materials Policy Commission reports (1952), and U. S. Bureau of Mines Bull. 556 (1956).

With regard to the question as to whether or not the United States is a representative sample of the earth's crust, the United States is a large area, about one-seventeenth of the land area of the world, exclusive of Antartica; it may not contain as large an exposed shield area as other areas of comparable size, but it does have all the major kinds of geologic terranes found anywhere, and may be accepted as a reasonably representative sample of the earth's crust. Although the discoveries still being made in the United States amply show that it is incompletely explored, it is reasonable to assume that the exploration completed has been intensive enough to indicate the relative magnitude of the reserves of many elements. As for the influence of the mineralogy of ore elements on their amenability to processing, technologic problems do make difficult the recovery of certain elements from some mineralogical types of

occurrences—nickel from nickel silicates, for example—but for most elements the important mineralogical types are susceptible to treatment; in other words, gold, silver, copper, and many other elements can be recovered from all their quantitatively important deposits that contain more than the cut-off grade. And most of the elements in the group studied are ones that either form distinct ores, or, like cadmium, are dispersed in ores of other elements from which they can be recovered as co-products.

Although all this applies to many elements in the United States—enough, aparently, to bring out the close correlation between reserves and abundance shown for several elements on figure 1—it does not apply to all. Aluminum is a good example of an element that shows a relatively large departure from the idealized regression lines of $R = A \times 10^9$ to 10^{10} on figure 1, conceivably because of the influence of some of the factors named above. Unlike many other metals, aluminum has been in demand for only a few decades, and the terrane of the United States has not been as thoroughly prospected for it as for most other metals. Moreover, aluminum cannot yet be recovered economically from most silicate minerals, so aluminum reserves include only the oxides. And these are formed mainly as the result of tropical or subtropical weathering processes that do not operate to an appreciable extent within the limits of the continental United States. The net effect of all of these factors is to yield smaller reserves of aluminum, compared to other elements, than otherwise might be expected for such an abundant element.

The scatter or variation in the reserve-abundance ratios for all the elements shown on figure 1 conceivably could be the result of factors similar to those that seem to yield the comparatively small aluminum reserves. In fact, examination of the distribution of elements on figure 1 shows that most of those that lie close to the $R = A \times 10^9$ regression line are ones that were sought and used in ancient times, mainly because their ores were easily smeltable. Those that show a rather large departure from this line are mostly ones relatively new to commerce and industry; they have been sought only in recent times and some of their mineralogical types of occurrences are still not workable.[1] Additional exploration and technological advances probably will increase the minable reserves of this group, enough perhaps to bring them into the range of $R = A \times 10^9$ to 10^{10}, observed for the other group.

The explanation of the linear relation between reserves and abundance that prevails over most of the abundance range seems to lie in a combination of compensating factors. The grade-frequency distribution of the elements, over the range of their concentration in the earth's crust, seems to be approximately log-normal in analyzed samples (Ahrens, 1957); the largest amount of each element occurs in rocks whose grade is not far from the average in the crust, and the amount in each grade-class decreases progressively to the maximum grade in ores, which may be a few to a few thousand times that of the crustal average. In the range of the higher concentrations in ores, tonnage increases

[1] Tin is a notable exception to this generalization. It may be that for this element, and perhaps some others, the United States is not a representative sample of the earth's crust, and that it simply does not contain a proportionate share of the world's minable deposits. Unless some geologic reason for this is evident, however (and it is not evident for tin), such a conclusion must be accepted cautiously.

exponentially as grade decreases (Lasky, 1950; see also the tonnage estimates for various elements in the President's Materials Policy Commission Report, v. 2, 1952). The amount in each grade-class varies from element to element, but the pattern of distribution is seen from the data on many elements to be generally similar in all.

Now, at concentration factors of 1, the relation between tonnage and abundance is linear. Because of the general similarity in the pattern of grade-frequency distributions among the elements, the relation between tonnage and abundance is probably nearly linear also when the concentration factors are nearly uniform from element to element. As shown in figure 2, the concentra-

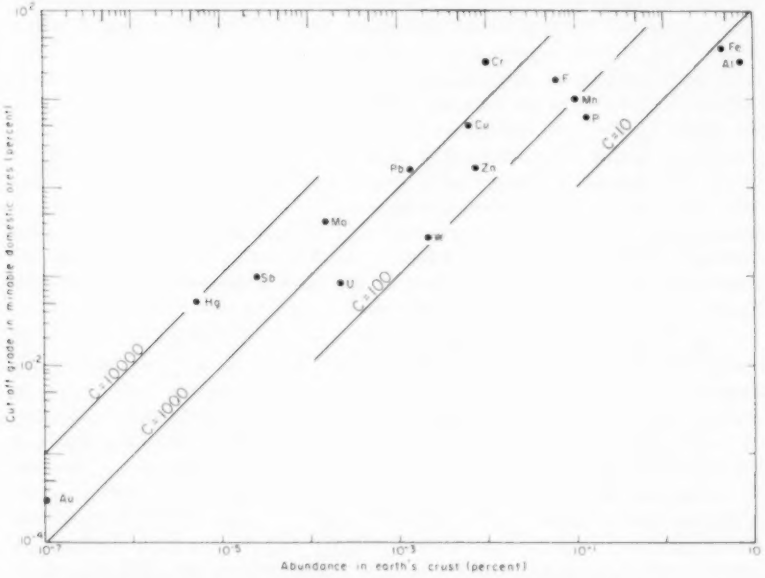

Fig. 2. Cut-off grade of minable ores compared to crustal abundance. Several concentration factors (C) over the observed range of cut-off grade and abundance are indicated by idealized regression lines.

tion factor at the cut-off grade of minable ores in the intermediate range of abundance varies from about 100 to about 1,000—a rather narrow range, compared to the variation in abundance and tonnage; and the concentration factors of minable ores of the abundant and rare elements are within about an order of magnitude of this range. The relatively small variation in the concentr[illegible]ctor among the moderately abundant elements seems to explain the li[illegible]relation between reserves and tonnage among them, so the problem is reduced to explaining the extension of this relation to the rare elements whose concentration factors are about an order of magnitude greater than those of intermediate abundance.

The cut-off grade for each element, of course, is partly fixed by mining and refining costs, and the good relation of cut-off grade to abundance shown in figure 2 suggests that costs are strongly influenced by abundance. This relation was found and stated in a somewhat different way by John R. Menke (quoted by Schurr and Marschak, 1950, p. 22), who found, by comparing the price of refined minerals with the grade in marginal ore bodies, that

> the cost of mining and refining *per lb. of ore* of widely differing minerals is remarkably alike. . . . In other words, then, the cost of minerals won by similar mining methods is determined in good part by *the number of lbs. of ore mined per lb. of pure mineral.* (italics Shurr and Marschak's)

Cost is also influenced by other factors—demand, ease of concentrating and refining the ore, and so on. What has apparently happened with rare elements such as gold and mercury is that exhaustion of high-grade ores, coupled with a combination of technology and ore mineralogy that yield relatively low recovery costs, have both forced and permitted the mining of ores one or two orders of magnitude lower in grade than the highest grade ores of these elements. This has made available an exponentially larger tonnage of ore—enough to yield reserves of a size more comparable to those of elements in the intermediate range. This has not yet happened to any appreciable extent to the more abundant elements, for their cut-off grades have not yet been much lowered; as advancing technology and exhaustion of high grade deposits permits and forces mining of lower grade ores, however, minable reserves of the abundant elements will increase to values closer to the ideal regression lines shown in figure 1 (see data there on low grade resources of iron and aluminum).

The linearity of the relation between reserves and abundance shown in figure 1 is thus almost fortuitous for the rare elements, and more data (and analysis of data already available) are needed on the relation between grade and tonnage for each element before we will be able to predict minable tonnages available at lower cut-offs. But the empirical relation shown in figure 1 has practical application now in estimating reserves available at a given cut-off in large areas or in the world as a whole, and also in the evaluation of estimates already made. These applications are indicated by the comparisons shown in table 1 between known world reserves and the minimum tonnages of minable ores that might be expected on two assumptions: one, that the rest of the world contains at least as much minable material in direct proportion to its size as is known in the United States; the other, that the world contains at least as much minable material in proportion to its size as is known or expected in the United States on the basis of $R = A \times 10^9$ to 10^{10}.

A simple extrapolation of the known reserves of the United States to the rest of the world makes no assumption that a relation exists between crustal abundance and reserves, but merely assumes that large segments of the continental crust are similar in the aggregate in their content of ore, and that the United States is a good sample both with regard to composition and to degree of exploration. This approach was suggested by Nolan (1950) as a basis for estimating potential resources in covered areas within belts broadly favorable for the occurrence of ore, and it is a sound method for this purpose. It cannot be applied to the world at large, however, for elements, such as germanium,

that have not been sought or explored. Applied to other elements it has the danger of underestimating reserves that have been inadequately explored or that for some reason are not well represented in minable quantities in the United States. And the assumption that large segments of the crust are homogeneous in their content of ores of various elements is challenged by the uneven distribution of known reserves over the world (see Carlisle, 1959, p. 350). For example, most the world's important tin, diamond, nickel, and molybdenum deposits are restricted to a few areas, and their distribution gives the impression that no segment of the crust is representative of the whole.

TABLE 1

Comparison of known and expectable minable tonnages of several elements in the world (short tons)

	A. Known reserves[1]	B. Tonnage forecast from known U. S. reserves × 17.3[2]	C. Tonnage forecast from $R=A\times10^9$ to $10^{10}\times17.3$[3]
Antimony	3.8×10^6	2.1×10^6	$0.52 - 5.2 \times 10^6$
Copper	2.3×10^8	4.3×10^8	$1.2 - 12 \times 10^8$
Iron	5.9×10^{10}	4.3×10^{10}	8.0×10^{10}
Gold	3.4×10^4	2.8×10^4	$0.2 - 2 \times 10^4$
Lead	4.5×10^7	14×10^7	$2.8 - 28 \times 10^7$
Mercury	1.4×10^5	3.8×10^5	$1 - 10 \times 10^5$
Tin	7.3×10^6	—	$7.8 - 78 \times 10^6$
Aluminum	6.4×10^9	2.1×10^9	1400×10^9
Chromium	2.3×10^8	0.4×10^8	$3.0 - 30 \times 10^8$
Fluorine	1.1×10^7	6.4×10^7	$110 - 1100 \times 10^7$
Manganese	4.9×10^8	0.3×10^8	$17 - 170 \times 10^8$
Thorium	9.2×10^5	7.8×10^5	$220 - 2200 \times 10^5$
Tungsten	1.3×10^6	1.4×10^6	$61 - 610 \times 10^6$
Uranium	8.8×10^5	33×10^5	$45 - 450 \times 10^5$
Zinc	8.0×10^7	36×10^7	$14 - 140 \times 10^7$

[1] Data on reserves of uranium and thorium from Bowie (1959); all other data from President's Materials Policy Commission (1952).

[2] The land area of the world (52,300,000 square miles excluding polar regions) is about 17.3 times that of the United States (3,022,000 square miles). Known U. S. reserves are extrapolated to the world in direct proportion to area of land surface.

[3] Assumes, from the relations shown in figure 1, that the United States contains minable reserves of all elements (except the most abundant) equal to their crustal abundance in percent times 10^9 or 10^{10}. The minable tonnage so derived is then extrapolated to the rest of the world as in column B. Because both figures 1 and 2 suggest that 10^{10} is a factor too high for the abundant elements, the estimates for iron and alumina are based only on a factor of 10^9.

Nevertheless, the assumption that the relatively unexplored part of the world contains tonnages proportionate to that in the known part is a promising means of estimating world reserves of the better known elements, and its efficacy is well shown by comparing columns A and B of table 1. The first group of elements in this table are those valued since ancient times; with the exception of tin and lead, the estimates prepared by this method for these elements

are extremely close to the known reserves. The estimates for the second group of elements, those sought and used only in recent times, do not compare as favorably with known reserves as those for the first group. With the exception of manganese, however, they are all within an order of magnitude of the estimates of known reserves. Lacking a better approach, estimates prepared by this method surely would be satisfactory, at least for the well explored elements in the known segment of the crust.

An estimate of the world reserves, or of any large segment of the world's land area, based on the assumption that minable tonnages of the elements are proportional to crustal abundance not only has the foundation carried by observations in a supposedly representative sample, but it has the added strength of theory to meet the challenge provided by the uneven distribution of known deposits—that is, it gives reason for believing that the distribution of known deposits in relatively unexplored parts of the world is more a function of chance discovery than of representative sampling. And it provides a basis for estimating world reserves of elements that have not been adequately explored in the United States or other repreesntative areas.

Column C of table 1 is based on this assumption. Comparison with column A shows that for the first group of metals, forecasts of world reserves prepared in this way bracket the estimates of known reserves for four of the elements, and their maxima are higher than the estimates of known reserves for all elements except gold (the most sought for element of all, the one whose reserves are perhaps best known, and the one whose cut-off grade has decreased the most over the history of its mining and use); in other words, estimates based on the maximum reserve-abundance ratio ($R = A \times 10^{10}$) observed in the United States exceed the estimates of known reserves by a margin that allows for additional discoveries.

Estimates prepared in this way for the second group are all higher than the estimates of known reserves and they suggest that this group of elements—relatively new to commerce—has been poorly explored indeed. This probability is somewhat counter-balanced, however, by the fact that these elements were not used in ancient times for the good reason that their mineralogical types were refractory, and some of them still are. If 1.4×10^{11} tons of aluminum exist in deposits of comparable grade to that in the known bauxite deposits, much of it probably is in silicate minerals not workable by available technology.

In spite of the danger that this method of estimating reserves in unknown areas is likely to over-estimate the tonnages available under present technologic and economic conditions, it seems the better of the two in that it allows for new discoveries, and permits the forecast of previously unexplored elements. Considering the uncertainties involved in the data used to prepare column C, the numbers shown in column C are unnecessarily refined. For purposes of estimating world reserves of unexplored elements, a figure of something like $A \times 10^{10}$ to 10^{11} probably will give the right order of magnitude.

Forecasts made either by extrapolating known reserves in a small area to a large one, or by extrapolating on the basis of a reserve-abundance ratio, yield minimal estimates. Both methods imply that reserves in the known area

are fully established and they assume that only some minimum thickness of the earth's crust is within the reach of exploration and mining; they also assume that the technologic and economic conditions that have established the present cut-offs will remain constant. Advances in technology will invalidate these assumptions and require new and larger estimates. Nevertheless, such forecasts are useful in indicating the order of magnitude of reserves that can be expected over the world or large segments of it with the technology available now. And it seems reasonable to expect that better knowledge of the relations between concentration factor, size of individual deposits, and tonnage of various elements will permit us to predict rather accurately how much ore of a given grade is contained in representative segments of the earth's crust.

ACKNOWLEDGMENTS

I am much indebted to P. W. Guild, M. R. Klepper, W. L. Newman, and W. G. Schlecht for reading and constructively criticizing this paper; and I am especially grateful to E. M. Shoemaker, who contributed much to the explanation presented here for the relation between reserves and abundance.

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CLASSIFICATION AND ASSOCIATION OF THE CARBONATE MINERALS OF THE GREEN RIVER FORMATION*

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ABSTRACT. About one fourth of all known carbonate minerals occur in the Green River formation of Wyoming, Utah, and Colorado, a sequence of Eocene lacustrine deposits. Most are alkali (sodium) or alkali-earth (calcium, magnesium, barium) or alkali-alkali earth carbonates; a few contain rare earths, or aluminum, chlorine, or phospate. Twenty well-defined species are listed, and arranged in accordance with Dana's System, 7th Edition. The particular assemblage found in each of the three States implies corresponding variation in the over-all geochemical and geological conditions affecting their deposition.

The Green River formation of Wyoming, Utah, and Colorado (map, fig. 1) is a sequence of essentially lacustrine deposits of early to middle Eocene age, whose unique characteristics were first systematically explored by Wilmot H. Bradley in a series of classical papers (Bradley 1926, 1930a, 1930b, 1931, 1936, 1948, 1959). This formation, now known to contain the world's largest reserve of hydrocarbons, in the form of oil shale, also contains a tremendous reserve of trona ($Na_2CO_3.NaHCO_3.2H_2O$), conservatively estimated at several

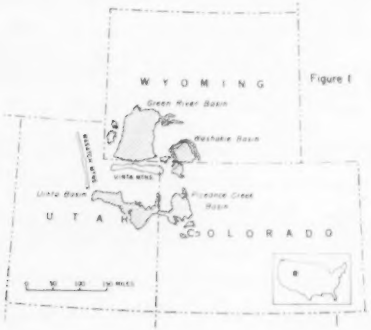

Figure 1

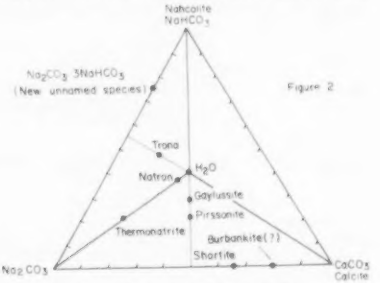

Figure 2

Green River minerals in the system $NaHCO_3 - Na_2CO_3 - CaCO_3 - H_2O$

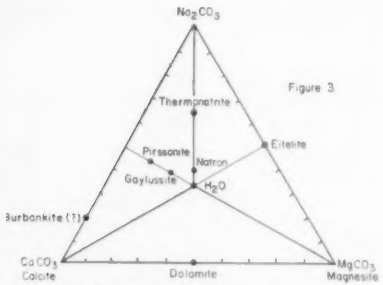

Figure 3

Green River minerals in the system $Na_2CO_3 - CaCO_3 - MgCO_3 - H_2O$

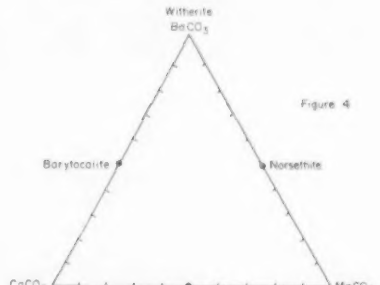

Figure 4

Green River minerals in the system $CaCO_3 - MgCO_3 - BaCO_3$

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billions of tons. Trona is readily converted to soda ash (sodium carbonate), a basic industrial commodity; and within a few years development of the Wyoming deposits will increasingly supersede soda ash manufacture by the Solvay process. The mineralogy of the Green River formation thus is a matter of great interest, in particular because of the remarkably varied carbonate minerals found—some twenty distinct species. Actually about a fourth of all carbonate species known occur in the Green River formation.

Because data on the Green River carbonates are widely scattered and much information is yet unpublished, it is thought that the synoptic review of their mineralogy here presented—necessarily skeletal though it be—will be helpful to the student of these matters. A similar summary discussion of the remarkable silicate mineralogy of the Green River formation—including twelve species—has been submitted for publication in the proceedings of the XXI International Geological Congress at Copenhagen in 1960.

In this limited paper, it will not be possible to give any detailed account of the mineralogy or geochemistry of individual species, or of their geological relationships to the Green River formation; for this the reader is referred to the literature listed in part in the references cited below.

Table 1 lists the carbonate minerals of the Green River formation, and indicates their occurrence in three western states. The tabulation follows the classification of Dana's System of Mineralogy, 7th edition, volume 2.

The distribution indicated under the heading Occurrences in table 1 appears to be much more than merely geographic or chance discovery; it is significant of specialization of geological history in Green River time in the three major sub-basins of the Eocene lakes, known as the Green River basin in southwestern Wyoming, the Uinta basin in northeastern Utah, and the Piceance Creek basin in northwestern Colorado. Milton and Eugster (1959) have attempted to relate the difference in sodium carbonate mineralogy of Wyoming (characterized by trona) and the other two states (characterized by nahcolite) to limnologic considerations; here we may say only that in each of the three states, so far as our present knowledge goes, there is a distinctive carbonate (and other) mineralogy.

Calcite and dolomite are the dominant minerals of the Green River formation. These, with quartz and organic matter, and lesser but significant amounts of pyrite, albitic feldspar, analcite, detrital mica, and clay minerals, constitute the major rock-forming minerals of the Green River beds in all three states. Doubtless calcite is in general more abundant than dolomite, although locally in some beds or series of beds dolomite may exceed calcite. In outcrop the once present sodium-containing carbonates are invariably replaced by calcite; this is particularly true of shortite. The so-called "oil shale" of the Green River is largely dolomite or limestone with varying organic content, as is also most of the "marlstone" of various writers.

Shortite is the most abundant of the sodium-containing carbonates in the Wyoming Green River beds, and followed by trona and northupite is the most abundant carbonate (other than calcite and dolomite) found in the John Hay

Jr. No. 1 well in Wyoming; study of the core from this well has recently been completed by J. J. Fahey, and the data submitted for publication as a professional paper of the U. S. Geological Survey. Much shortite, but no trona, is also found in the Uinta basin in Utah; neither mineral has been found in the Piceance basin in Colorado. Nahcolite, abundant in Utah and Colorado, is found only as traces in Wyoming.[1]

Microscopically disseminated bartytocalcite is found rather widely but never abundantly in Utah and Wyoming; accompanied by burbankite, as a microscopic rarity. In Utah, barytocalcite is found with eitelite and a unique asemblage of silicate minerals. In a well core from Colorado, dawsonite occurs through several hundred feet; cryolite is present also.

In Wyoming, all of the carbonates occur in a series of calcareous or dolomitic beds containing bedded trona. Also in Wyoming, and only there, in addition to shortite, trona, and northupite, the following have been identified: pirssonite, gaylussite, bradleyite, witherite, thermonatrite, barytocalcite, nordsethite, natron, and a new unnamed species with the composition Na_2CO_3. $3NaHCO_3$. Trona, thermonatrite, natron, and the new $Na_2CO_3.3NaHCO_3$ are all phases in the Na_2O-CO_2-H_2O system; and similarly shortite, pirssonite, and gaylussite are phases of the more complex Na_2O-CaO-CO_2-H_2O system. The phase relationships in these comparatively simple systems are summarized by Milton and Eugester (1959). Carbonate minerals known to exist in these systems, and in the $CaCO_3$-MgO_3-$BaCO_3$ system are shown in figures 2, 3, and 4. The chlorine in locally abundant northupite, and the phosphate in bradleyite indicate that the actual physical chemistry of the trona deposits involves a good deal more than these three simple systems. It may also be mentioned in this connection that study of these carbonates, e.g. trona and northupite, has shown remarkable intergrowths with silicates. Thus trona crystals enclose loughlinite $Na_2O.3MgO.6SiO_2.8H_2O$ (Fahey, Ross, and Axelrod, in press) and labuntsovite (K,Ba,Na,Ca,Mn) (Ti,Nb) $(Si,Al)_2(O,OH)_7H_2O$ (Milton, Mrose, Fahey, and Chao, 1958); and northupite encloses magnesioriebeckite $Na_2(Mg,Fe)_3(Fe,Al)_2Si_8O_{22}(OH)_2$. The relations suggest simultaneous crystallization of carbonate and silicate.

Brief mention may be made of the other minerals of table 1. Magnesite has been noted but once, in an outcrop sample from Utah. It forms good crystals several millimeters long in a dolomite matrix. Siderite is probably more abundant than can now be shown; it has been noted in microscopic spheroids in well cuttings from Utah, as yellow microscopic rhombohedra in a well core from Colorado and in an outcrop sample from Wyoming.

To be considered as a phosphate rather than a carbonate mineral is carbonate fluor-apatite (Love and Milton, 1959). This mineral constitutes a significant portion of certain beds in the Green River of Wyoming, some of which have been traced continuously over scores of miles. Almost invariably it carries traces of uranium. The mineral is cryptocrystalline and invariably submicroscopically intergrown with dolomite, calcite, analcite, and quartz.

[1] Since this was written, recent drilling has exposed massive nahcolite with trona and halite in Wyoming.

TABLE 1

Green River carbonates

(Arranged according to Dana's System, 7th ed., v. 2, p. 132-133)

	Occurrence		
		Utah, Colorado	
Class 13 Acid carbonates Type 1 Miscellaneous			
13.1.1 Nahcolite $NaHCO_3$	Wyoming	"	
13.1.4 Trona $Na_3CO_3.NaHCO_3.2H_2O$	"		
*13.1.5 New species, unnamed[1] $Na_2CO_3.3NaHCO_3$			
Class 14 Anhydrous normal carbonates Type 1 $A(XO_3)$			
14.1.1 Calcite group			
14.1.1.1 Calcite $CaCO_3$	"	"	"
14.11.2 Magnesite $MgCO_3$			
14.11.3 Siderite $FeCO_3$	"	"	"
14.1.3 Aragonite group			
14.1.3.2 Witherite $BaCO_3$	"		
Type 2 $AB(XO_3)_2$			
14.2.1 Dolomite group			
14.2.1.1 Dolomite $CaMg(CO_3)_2$	"	"	"
14.2.3 Barytocalcite $CaBa(CO_3)_2$	"	"	
*14.2.4 Norsethite[2] $MgBa(CO_3)_2$	"		
Type 3 $A_2B(XO_3)_2$			
*14.3.1.2 Eitelite[3] $Na_2Mg(CO_3)_2$		"	
Type 4 $A_2B_2(XO_3)_3$			
14.4.1 Shortite $Na_2Ca_2(CO_3)_3$	"	"	
Type 5 $A_2B_4(XO_3)_5$			
*14.5.1 Burbankite[4] $Na_2(Ca,Ba,Sr,La,Ce)_4(CO_3)_5$	"	"	
Class 15 Hydrated normal carbonates Type 1 $A(XO_3).xH_2O$			
15.1.1 Thermonatrite $Na_2CO_3.H_2O$	"		
15.1.6 Natron $Na_2CO_3.10H_2O$	"		
Type 2 Miscellaneous			
15.2.2 Pirssonite $Na_2Ca(CO_3)_2.2H_2O$	"		
15.2.3 Gaylussite $Na_2Ca(CO_3)_2.5H_2O$	"		
Class 16 Carbonates containing hydroxyl or halogen			
16.2.1 Dawsonite $NaAl(CO_3)(OH)_2$			"
16.2.2 Northupite $Na_3Mg(CO_3)_2Cl$	"		
Class 17 Compound carbonates			
17.1.2 Bradleyite $Na_3Mg(CO_3)(PO_4)$	"		

* References to table 1

[1] New species, unnamed $Na_2CO_3.3NaHCO_3$ unpublished data.

[2] Norsethite, see Milton, Mrose, Chao and Fahey, 1959.

[3] Eitelite, see Milton, Axelrod, and Grimaldi, 1954.

[4] Burbankite, see Pecora and Kerr, 1953; and Milton and Eugster, 1959.

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A NEW METHOD OF PLOTTING CHEMICAL ANALYSES OF BASALTIC ROCKS*

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ABSTRACT. In studying the differentiation of basaltic magmas by means of the chemical analyses of basaltic rocks, very useful diagrams are obtained by plotting percent CaO, MgO, or Na_2O+K_2O against the Al_2O_3/SiO_2 weight ratio. The diagrams show the tholeiitic and the alkalic series of basaltic rocks nicely separated and parallel in their variations. Fractional crystallization of clinopyroxene appears to be the principal mechanism by which tholeiitic magmas are converted to alkalic magmas.

INTRODUCTION

Many different ways of plotting chemical analyses of basaltic rocks have been tried by previous investigators to represent and interpret the trends of differentiation of basaltic magmas. In recent years, less use has been made of conventional variation diagrams and more specialized plots of selected oxides have been favored. Among these are the plots of total alkalies vs. SiO_2 by Tilley (1950) and of MgO vs. SiO_2 by Powers (1955). Other recent chemical diagrams of basaltic rocks include the type published by Wager (1956), in which the plagioclase composition is plotted against the ratio of iron to mafic elements, and the one by Kuno and others (1957) in which the individual oxides are plotted against a solidification index, a ratio of MgO to several other oxides.

Each method of plotting has illuminated one or more facets of differentiation of basaltic magmas, but none has shown in a satisfactory manner a possible genetic relationship between the two major series of basaltic rocks, the tholeiitic and the alkalic. Also, although basaltic rocks consist essentially of only three major minerals (olivine, clinopyroxene, and plagioclase feldspar) none of the diagrams, except that of Powers (1955), has directly related the compositional variations to variations in the proportions of the major minerals. Powers' diagram (MgO vs. SiO_2) clearly indicates the manner in which the composition of most Mauna Loa and Kilauea lavas is controlled by the amount of olivine they happen to contain.

A new method of plotting the composition of basaltic rocks has been devised which seems to offer major advantages heretofore not realized. The new plots, which use the Al_2O_3/SiO_2 weight ratio as the abscissa, were developed during the course of a comprehensive study of the geochemistry and geophysics of basaltic volcanism at the U. S. Geological Survey's Hawaiian Volcano Observatory. The full results of this study will be published elsewhere by D. H. Richter, J. P. Eaton, and the writer, and will emphasize the relationship between the speed of ascent of the magma and the composition of the erupted lava. This paper is restricted to outlining a method of plotting the chemical data and does not discuss magmatic differentiation in any great detail.

SOURCES OF DATA

No one region of the earth's surface seems to provide the complete suite of

* Publication authorized by the Director, U. S. Geological Survey.

tholeiitic and alkalic basalts. Therefore, it has been necessary to utilize analyses of rocks from both Hawaii and from the Tertiary Hebridean basaltic province of Great Britain in our study. The 150 analyzed samples from these two areas seem to represent fairly adequately the range of composition and mineralogy of tholeiitic and alkalic basalts.

Analyses of Hawaiian rocks were compiled from publications by Cross (1915), Wentworth and Winchell (1947), Macdonald (1949, 1955), Macdonald and Eaton (1955), Muir, Tilley, and Scoon (1957), and Kuno and others (1957). Those of the Hebridean rocks are from the classic memoirs on Mull (Bailey and others, 1924) and on Ardnamurchan (Richey and others, 1930).

The compositions of the major minerals with respect to several elements can also be plotted on the new diagrams. Although the compositional variation within the plagioclase series is adequately indicated by the Al_2O_3/SiO_2 ratio, the variations in clinopyroxene and olivine cannot be represented without unduly complicating the diagrams. Therefore, an average composition with respect to the plotted constituents is used as a reference point for these two minerals. Analyses of clinopyroxenes and olivines were assembled from several sources (Washington and Merwin, 1922; Macdonald, 1949; Ross, Foster, and Myers, 1954; and Muir, Tilley, and Scoon, 1957).

Plots of percent CaO and $Na_2O + K_2O$ vs. Al_2O_3/SiO_2 weight ratio for 150 Hawaiian and Hebridean rocks are shown in figures 1a and 1b respectively. Figure 2a is a plot of MgO vs. the same ratio. Figure 2b is a skeletonized version of figure 2a and shows the postulated paths of differentiation of basaltic magmas.

The great advantage offered by the indicated compositions of the three major minerals of basaltic rocks, olivine, clinopyroxene, and plagioclase, is immediately apparent. Differences in bulk chemical composition are readily interpretable in terms of differences in the relative proportions of the three minerals. The plots, furthermore, may be compared directly with existing phase diagrams involving basalt-like compositions and allow the full force of phase-equilibria data to be brought to bear on the problem. A somewhat more accurate comparison would be possible if the accessory minerals, such as ilmenite, magnetite, and apatite, were subtracted and the individual rock analyses recalculated to 100 per cent. However, this refinement is not considered necessary in the present study in which only the main trends of differentiation are being sought.

TRENDS OF DIFFERENTIATION

Despite a scattering of points, certain definite trends in composition are evident in both figures 1 and 2, and these trends have been delineated by means of heavy lines. The scatter of points must be due in large part to the above-mentioned averaging of the compositions of the mafic minerals, to the non-uniform amounts of auxiliary minerals, and to errors in chemical analysis. The effect of analytical error can be large because alumina is one of the least accurately determined constituents in a conventional rock anaylsis.

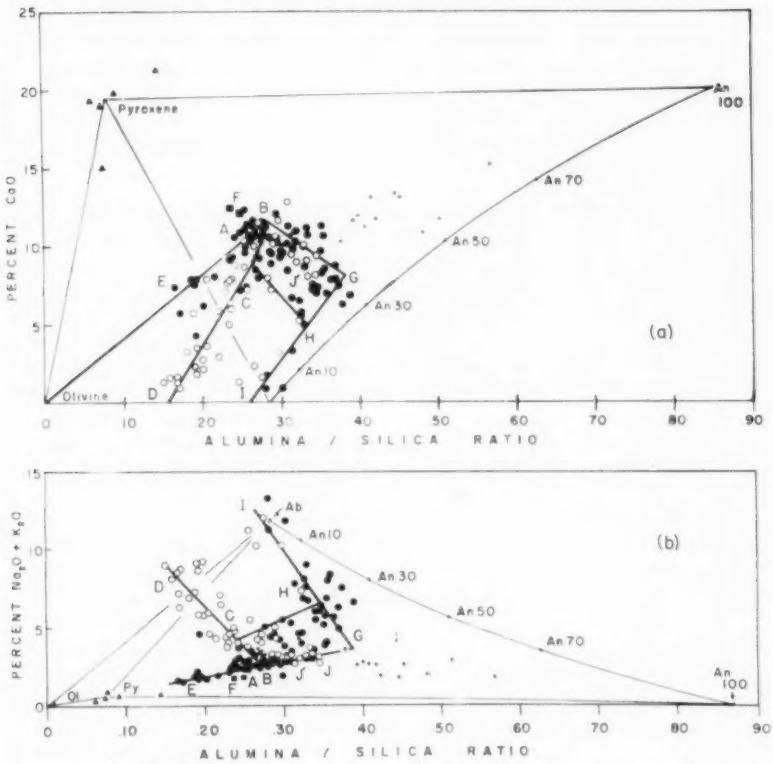

Fig. 1. Trends in bulk composition of Hawaiian rocks (black circles) and of British Hebridean rocks (open circles) as indicated by a plot of CaO against the Al_2O_3/SiO_2 weight ratio (diagram a), and of Na_2O+K_2O against the same ratio (diagram b). Types of rocks represented by the letters A to J′ are listed in the text.

Small black triangles indicate the composition of Hawaiian clinopyroxenes. Black squares at F mark the composition of Hawaiian ankaramites. Crosses represent plagioclase-rich rocks (central porphyritic type) of the Hebridean province. Small dots on the line between B and C represent a series of lavas collected during the 1955 eruption of Kilauea.

Figure 1 is awkward because quartz ($Al_2O_3/SiO_2=0$) and olivine plot close together near the origin in both diagrams and because the composition of alkalic olivine basalt does not show clearly in diagram (a). But the figure has been included to show that CaO and Na_2O+K_2O indicate essentially the same trends of differentiation of the rocks as does MgO, when plotted against the Al_2O_3/SiO_2 ratio. Listed below are the rock types corresponding to the lettered points in the different parts of the diagrams.

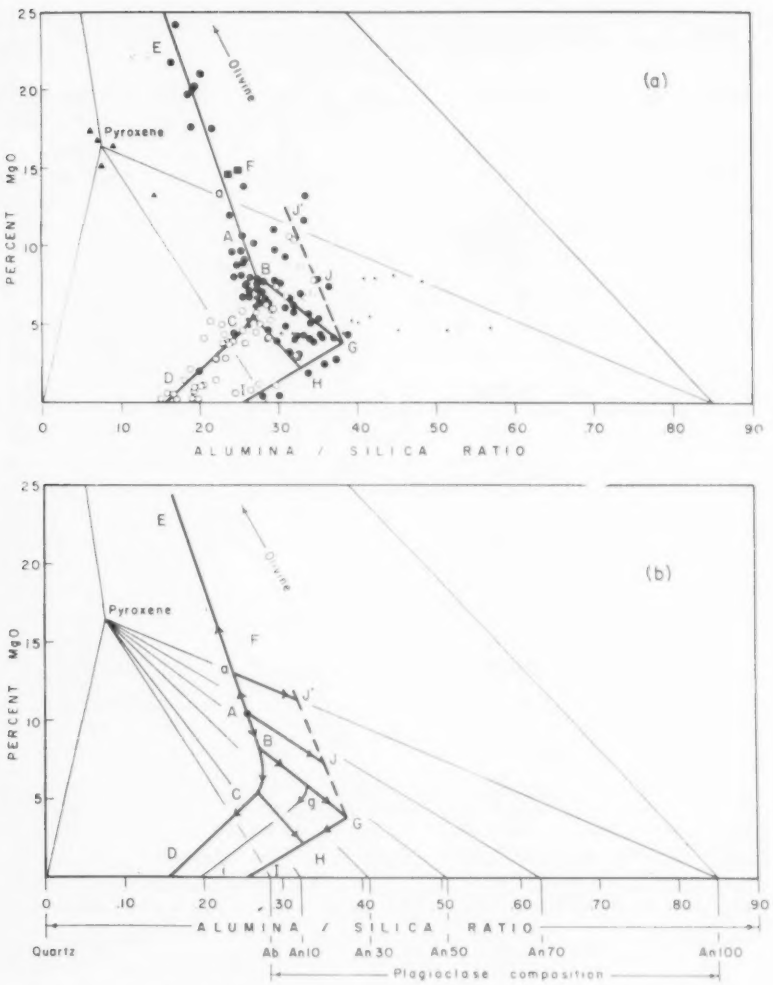

Fig. 2. Trends in bulk composition of Hawaiian and British Hebridean rocks as indicated by a plot of MgO against the Al_2O_3/SiO_2 weight ratio (diagram a). The symbols are the same as in fig. 1. Average Hawaiian olivine, which plots off the diagrams, has the following composition: MgO, 47.2%; CaO, 0.25%; Al_2O_3/SiO_2, 0.009. Diagram (b) is a skeletonized version of diagram (a), and shows changes in composition arising from fractional crystallization of one or more minerals. The tholeiitic series of basalts lie along E-a-A-B-C-D, and the alkalic series, along J'-J-G-H-I. Rocks representing trends intermediate between these two major series lie along g-i.

Note the absence of Hebridean rocks with composition along E-a-A—B, and the absence of Hawaiian rocks corresponding to the plagioclase-rich rocks of the Hebrides. Hawaiian rocks with compositions along C-D are scarce.

Tholeiitic basalt series

A. Tholeiitic olivine basalt
B. Tholeiitic basalt
C. Quartz basalt
D. Granophyre
E. Picrite (oceanite)

Alkalic basalt series

F. Ankaramite
G. Hawaiite (andesine andesite)
H. Mugearite (oligoclase andesite)
I. Trachyte
J. Alkalic basalt
J′. Alkalic olivine basalt

Figure 2 most clearly represents the relationships among the different types of rocks and further discussion will be based mainly on it. A single fundamental tholeiitic magma (A) is believed to produce all members of the tholeiitic and alkalic series through fractional crystallization. As has been noted by previous investigators, the composition of such a fundamental magma corresponds essentially to a mixture of clinopyroxene and plagioclase feldspar (An≈60). It also closely approximates the average composition of lavas from the currently active Hawaiian volcanoes, Mauna Loa and Kilauea (Macdonald, 1949).

The parallelism in composition between the tholeiitic basalt series (E–a–A–B–C–D) and the alkalic basalt series (J′–J–G–H–I) is well shown in figure 2. Both series have olivine-rich members (E–a–A and J′–J) and a group of closely related differentiates with progressively increasing content of silica (B–C–D and G–H–I). In the tholeiitic series, this group includes rocks such as granophyre (D) that are rich in quartz whereas in the alkalic series even the most siliceous members (I≈62% SiO_2) are free of quartz or have only minor amounts of it.

Line E–a–A–B clearly shows the effect of early fractional crystallization of olivine and thus corresponds to Powers' olivine-control line (1955). Compositions resulting from the removal of olivine from the fundamental magma lie along A to B, while those due to the addition of olivine to the subjacent magma lie along A–a–E.

The parallel trend (B–C–D and G–H–I) in the two series is mainly the result of fractional syncrystallization of clinopyroxene and plagioclase feldspar. The progressive change in the composition of the plagioclase toward albite along this path of differentiation is nicely shown. Figure 1(b) also clearly indicates the increasing proportion of alkalic feldspar along this path in the two series.

The alignment of points in figure 2a strongly suggests that fractional crystallization of clinopyroxine is the general mechanism for converting tholeiitic magmas into alkalic magmas. In figure 2b the paths of such a differentiation involving fractional crystallization of clinopyroxene are indicated starting from four illustrative points (a, A, B, and C) in the tholeiitic series. There are differences in the details of the fractional crystallization process and in the nature of the differentiation product along the four paths but a discussion of these differences will be deferred to a subsequent paper.

It is highly significant that large phenocrysts of clinopyroxene (up to 2 cm in length) are found in greatest abundance in Hawaii in a type of lava known as ankaramite. As was shown by Macdonald (1949) for Mauna Kea

and other volcanoes, ankaramitic lavas occur among and overlie the last outpourings of tholeiitic lavas and are, in turn, overlain by alkalic basalts that are free or nearly free of clinopyroxene phenocrysts. Thus ankaramite, rich in coarse phenocrysts of clinopyroxene, constitutes a transitional type between the two basalt series, and focuses the attention on the role of clinopyroxene in the conversion of tholeiitic magma into alkalic magma.

If settling of clinopyroxene is the key process in converting magmas of the tholeiitic series into those of the alkalic series, a question immediately arises as to the possible existence of series intermediate between these two. In figure 2, for example, fractional crystallization of clinopyroxene along B–G may conceivably be interrupted at any point on the line and be succeeded by fractional syncrystallization of clinopyroxene and plagioclase feldspar. Such a course of differentiation might be along B–g–i in figure 2b.

Many of the basaltic lavas from Easter Island (Bandy, 1937) and members of the tonalite-quartz monzonite series of Ardnamurchan (Richey and others, 1930) do plot along g–i and thereby support the idea that intermediate series are possible. The nomenclature of such series intermediate between the tholeiitic and the alkalic presents new problems.

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CARBON 13 IN LAKE WATERS, AND ITS POSSIBLE BEARING ON PALEOLIMNOLOGY

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ABSTRACT. Surface waters of dimictic lakes in Connecticut have an average δC^{13} of $-8.6‰$, but the hypolimnion is more depleted ($\delta C^{13} \geqslant -22.6‰$). Seston, mud, and mollusc flesh average ca. $-30‰$, but submerged pondweeds are richer. Four sources, partly evaluable from C^{13} ratios, can contribute to hypolimnetic CO_2: *a*, surface CO_2 ($\delta C^{13} \sim -9‰$), inherited from circulation; *b*, respiratory CO_2 ($\delta C^{13} \sim -30‰$) from oxidation of seston; *c*, fermentation CO_2 ($\delta C^{13} \sim -5‰$), here shown to be enriched at the expense of CH_4 ($\delta C^{13} \sim -70‰$); *d*, carbonate CO_2 ($\delta C^{13} \sim 0‰$), important in hard waters. Metabolism $= b + c$; *c* and *d*, both heavy, can be separated by C^{14} assay, but this work is incomplete. In Quassapaug (softest), where *d* is negligible, *c* represents 4 to 20 percent of total metabolism. In Queechy (hardest), *d* could account for all heavy CO_2. In Linsley Pond (medium-hard), after allowance for *d*, c appears to provide 40 percent of total metabolism. Discovery that fermentation CO_2 is heavy raises important questions about groundwater. Applied to paleolimnology, C^{13} assay may be ambiguous for the same reason: fermentation tends to raise ratios lowered by aerobic metabolism. In one meromictic lake in a humid region, permanent stagnation has not produced fractionation, presumably because of *c*; owing to dominance of *d*, a meromictic saline lake also fails to show fractionation.

The earliest organic deposits of the last Pleistocene Searles Lake are enriched in C^{13}, presumably because the lake redissolved much carbonate, and the C^{13} ratio fell only gradually. In the Middle Huronian iron-banded cherts of Michigan the $FeCO_3$-rich layers may be poorer in C^{13} than the SiO_2-rich layers, and the depletion of these siderites supports the lacustrine hypothesis.

INTRODUCTION

The studies reported here were prompted by the discovery, predicted by Iversen (private communication to E. S. Deevey, October 5, 1949) and discussed by Godwin (1951) and Bartlett (1951), that aquatic plants, in utilizing bicarbonate from ancient limestone as a source of photosynthetic carbon, can be markedly depleted in C^{14} as compared with terrestrial organic carbon. Radiocarbon analyses by the solid-carbon method (Deevey, et al., 1954) showed that animal and plant material from the marl-forming hard waters of Queechy Lake was so poor in the heavier isotope that fictitious ages as great as 2000 yr could be calculated from contemporary assays. The highly interesting limnological problems thus exposed could not be considered further without better knowledge of the lacustrine distribution of the much more abundant stable isotope C^{13}. A comprehensive investigation of both isotopes in the waters of the Lahontan and Bonneville basins has been published by Broecker and Walton (1959), but these lakes are so unlike the bodies of water ordinarily studied by limnologists as to leave many questions unanswered. Acquisition of a mass spectrometer by the Yale Department of Geology, and the appointment of one of us (Oana) as Research Fellow in the Geochronometric Laboratory, provided the opportunity to launch the investigation.

It is a particular pleasure to be able to dedicate this paper to the acknowledged master of paleolimnology, Wilmot Hyde Bradley. We regret as much as Dr. Bradley will that there has been no opportunity as yet to apply our techniques to Eocene Lake Uinta; but although our results are unsatisfactory in many respects, we count on the interest of a distinguished geologist whose boyhood was spent near Linsley Pond.

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METHODS

CO_2 for isotopic analysis is prepared by methods similar to those of Craig (1953): *organic matter* (20 mg) is burned in CO_2-free oxygen in a combustion tube, after removal of carbonates by washing with HCl; *carbonates* (100 mg well ground) are acidified with 50% phosphoric acid (which is less likely to attack organic compounds than 100% phosphoric acid); for *water*, a 1-liter sample is acidified with sulfuric acid, containing cupric ions to precipitate any H_2S, and the CO_2 is swept out by bubbling CO_2-free air through a glass filter; *gas* samples containing CH_4 are combusted in a furnace containing CuO at 850° C, after CO_2 is removed by condensation in a trap cooled with liquid nitrogen. The CO_2 resulting from these steps is then purified further by drying (repeated sublimation from one liquid nitrogen trap to another through an intermediate trap cooled by dry ice in acetone) and passed through a furnace containing copper at 450°-500° C to remove oxides of nitrogen. N_2O is the least tolerable impurity, since it has the same molecular weight as CO_2. The overall pressure in the system when CO_2 is fixed is about 3 mm Hg; use of a trap of known volume permits the total CO_2 content of a water sample to be determined manometrically.

The CO_2 is removed from the vacuum system in a sample flask and introduced into a mass spectrometer of double collector type, Consolidated Model 21-401. The ratio of peak heights of molecular weights 45 and 44 is measured in comparison with N. B. S. reference samples no. 20 (Solenhofen limestone) and no. 21 (spectroscopic graphite). The C^{13}/C^{12} difference from Craig's standard, PDB-Chicago, is calculated from the ratios given by Craig (1957): PDB-Chicago, 0.0112372; N. B. S. 20, 0.0112253; N. B. S. 21, 0.0109249. The ratios are reported [2] as δC^{13} in per mil (‰) difference from PDB-Chicago, and are considered to be accurate within ± 0.3 per mil.

The new C^{14} analyses quoted in this paper were made by converting the sample to CO_2, purified by the methods introduced by DeVries and Barendsen (Barendsen, et al., 1957), and comparing the activity observed during several 20-hour periods with that of CO_2 from anthracite and from modern wood (hemlock, tree-ring-dated A.D. 1840-1850). The activity is extrapolated to zero

[2] $$\delta C^{13} = \frac{C^{13}/C^{12}_{(\text{sample})} - C^{13}/C^{12}_{(\text{standard})}}{C^{13}/C^{12}_{(\text{standard})}} \times 1000$$

Most values are negative, implying depletion in the heavier isotope, because the standard (marine carbonate) is particularly rich.

age, thus correcting for the Suess effect, but measurements of interlaboratory standards and of post-1954 modern samples recording artificial C^{14} are incomplete. Lake-water samples, ranging from 60 to 300 liters, are collected in 20-liter carboys, and the CO_2 evolved on acidification is swept to the vacuum line by bubbling CO_2-free air through a glass filter; removal of CO_2 from each carboy is usually complete within six hours, but may take several days if the CO_2 content is small.

Standard limnological methods are employed for the lake-water analyses: dissolved oxygen by the Alsterberg-Winkler method, alkalinity by titration to chlor-phenol red, acidity by titration to phenolphthalein. Temperatures are measured by means of a thermistor (Yellow Springs Instrument Co.) accurate to about $\pm$ 0.1°C. Gases are collected from mud in deep water by jiggling of a weighted funnel, 15 cm in diameter, fitted at the top with an inverted 500-ml bottle; the bottle is filled with lake water at the surface and the apparatus held upright until the gas has displaced the water and a rubber stopper can be inserted in place of the funnel.

THE CARBON CYCLE IN DIMICTIC LAKES

General statement

The reference standard for C^{13} content is a marine carbonate, and marine limestones and shells therefore average close to zero ($\delta C^{13} = -0.2‰$; Craig, 1953). The CO_2 of air is slightly depleted: –7.9‰ (Broecker and Olson, 1959). Seawater, supposedly at equilibrium with air, has a ratio very close to carbonate (Fonselius and Östlund, 1959). Surface lake water, at least in a humid region where carbonate rocks are absent, averages –8.6 per mil in our four softest lakes. The similarity of this figure to that of air is misleading; equilibrium with air should lead to an enrichment of about 5 per mil in the bicarbonate (Vogel, 1959), and the observed depletion is attributed to metabolism, not only in the lakes (as will be shown), but probably in circulating ground water. Whatever the case for surface waters, the deeper waters of stratified lakes are a special environment, and the hypolimnion is more like a plant or animal than it is like an equilibrium mixture of gaseous and ionic carbon.

It is well known that more CO_2 appears in the hypolimnion during summer than can be accounted for by the oxidation of seston (Ohle, 1934, 1952; Hutchinson, 1957, gives a general treatment). The "metabolic CO_2 anomaly" is computed by adding (1) free CO_2, (2) the CO_2 of ammonium bicarbonate, (3) half the CO_2 of alkali and alkaline earth bicarbonate, and subtracting (4) the CO_2 equivalent of the O_2 consumed in respiration. Most of this anomaly is reasonably attributed to fermentation of bottom deposits. In a lake as small as Linsley Pond, where the ratio of mud surface to water volume is high, and where density currents are known to transfer bicarbonate and reducing substances from shallow to deeper water, Hutchinson calculated the contribution of the mud, or pelometabolism, to be about equal to the more ordinary hydrometabolism. However, there is another, less obvious source of anomalous CO_2: particulate, perhaps colloidal, $CaCO_3$, precipitated mainly in the littoral zone by the photosynthesis of pondweeds, but redissolving in the more acid hypolimnion. Such a process has been invoked by Ohle (1952), working on the

very hard Grosser Plöner See, but would escape notice in the medium-hard Linsley Pond.

The isotopes of carbon serve as natural tracers, permitting quantitative identification of the sources of carbon in the biosphere. Given the understanding of lake metabolism that isotopic studies can provide, there are important applications to paleolimnology. As will be seen, the necessary understanding of metabolism is incomplete, and the applications are premature, since we restrict ourselves in this paper to the stable isotopes, C^{12} and C^{13}, and refer only incidentally to C^{14}.

The model: five New England lakes

The primary analytic data are summarized in tables 1 through 5. Most δC^{13} analyses entered are averages of duplicates, particularly inhomogeneous results being noted separately. The five lakes were chosen to provide maximal ranges, among the southern New England lakes that have been studied (Deevey, 1940), of the two properties, hardness and size. Linsley Pond is intermediate in hardness, but although it is the smallest in area its hypolimnion is relatively larger, compared to the epilimnion, than that of Queechy, the shallowest, or of Wononscopomuc, the deepest and largest. Quassapaug is large, deep, and very soft; Quonnipaug is small, as deep as Linsley, but softer; Queechy and Wononscopomuc are surrounded by limestone and very hard.

Noting that the bicarbonate of the surface waters has an isotopic ratio close to –9 per mil (where carbonate is absent), it is not difficult to account for the fact that net plankton (mainly algae in Linsley Pond, but mainly crustaceans in the other lakes) averages close to –30 per mil; this is the usual fractionation of carbon-fixation, and terrestrial organic carbon is depleted with respect to its source, the air, by almost exactly the same amount. Organic mud in deep water, whether formed mainly from plankton, from terrestrial plants, or from pondweeds, would be expected to have the same isotopic composition, as it does; but the pondweeds themselves present a separate problem considered later. The CO_2 in deep water, if it is formed *in situ* by oxidation of plankton and other organic seston in the presence of oxygen, would also be expected to have the same ratio, about –30 per mil; however, the values observed range from –22.6 per mil for Quassapaug to –10.2 per mil for Wononscopomuc, the range for Linsley Pond, from –18 to –11.9 per mil, being typically intermediate. The problem, then, is to explain these high values, and the extraordinary range among them.

A value of –30 per mil would not be expected, of course, for the *total* hypolimnetic CO_2: any CO_2 formed with that ratio from *oxidation* of seston would mix with *residual* CO_2, inherited from the last period of complete vertical circulation, and presumably having the ratio characteristic of surface waters. Quassapaug, with its value of –22.6, might be an example of such a two-component mixture. But two components alone will not account for the other results. Since the proportions of total CO_2 in epilimnion and hypolimnion are about the same in all the lakes there is no evidence of larger residues of surface CO_2 needed to raise the ratio from –22.6 toward –10 per mil. Moreover, the hypolimnetic δC^{13} values are already low early in the summer, when

residual CO_2 should be relatively more abundant, and they tend to *rise* with the progress of stagnation, as the oxygen is reduced nearly to zero and oxidative CO_2 is added. We were therefore not surprised to find that the CO_2 of *fermentation* gas, collected from the bottom deposits, can have a ratio as high as –4 per mil; the fractionation that is involved, leading in the opposite direction to values as low as –77.8 per mil for the methane, is discussed in a later section.

Qualitative features of this explanation are preesnted in the form of a diagrammatic lake (fig. 1), placed along a scale of δC^{13} values. It is somewhat idealized by rounding decimal places, and exaggerates the relative thickness of the hypolimnion, but is based on Linsley Pond in late summer. At least three components are shown as contributing to the hypolimnetic CO_2: residual or *surface* CO_2, with δC^{13} of –9 per mil; oxidative or *hydrometabolic* CO_2, with a ratio of –30 per mil; and fermentative or pelometabolic CO_2, with a ratio of –5 per mil. The resulting mixture has a ratio of about –12 per mil. The length of the line labelled *resultant* measures the relative importance of pelometabolism, being shortest in Quasapaug (mixture –22.6‰), longer in Quonnipaug (mixture –17.9‰), and longest in Linsley (mixture –11.9‰).

As stated, this explanation disregards a fourth source of CO_2 in deep water, the product of redissolution of carbonate rock, having an isotopic ratio near zero. The data from the hard-water lakes Queechy and Wononscopomuc show that this source cannot be disregarded there; both surface and deep waters are significantly enriched in C^{13}. Carbonate is negligible in Quassapaug, but may not be unimportant in the medium-hard lakes Quonnipaug and Linsley. To be safe we have included carbonate as a potential source of heavy carbon in the diagram.

Since we know the total amounts of CO_2 in surface and bottom waters of all the lakes, it is possible on certain assumptions to calculate the proportions of different components that contribute to an observed hypolimnetic ratio. The proportion of heavy CO_2 that appears in the hypolimnion in late summer is estimated (table 6) from two sets of analytic data: gasometric data for total CO_2 in the bottom water, which are reliable for that layer but overestimate the CO_2 in the whole hypolimnion, and titration data for the whole hypolimnion, which generally underestimate the total CO_2 owing to the arbitrary choice of end-point. Residual CO_2 is taken to equal that observed in surface water or in the whole epilimnion in early summer, with $\delta C^{13} = -8.3$ per mil, and the increment over this amount that appears in the bottom water or in the whole hypolimnion by the end of summer is considered to be metabolic CO_2, the sum of hydrometabolic CO_2 with $\delta C^{13} = -29.4$ per mil and pelometabolic CO_2 with $\delta C^{13} = -5.6$ per mil. The heavy CO_2 is shown as a percentage of the metabolic CO_2 in column 4, and as a percentage of the total deep-water CO_2 in column 5.

The results are only partly credible. It is not difficult to believe that pelometabolic CO_2 in a lake like Quassapaug might account for 4 to 20 percent of the metabolic, or 3 to 18 percent of the total deep-water CO_2. Quonnipaug has a slightly larger or slightly smaller pelometabolic contribution, depending on which set of assumptions is used, and this result is also reasonable. On the other hand, the basic assumption—that heavy CO_2 is all pelometabolic—fails for the two hard-water lakes, leading to values of the order of 60 percent of the

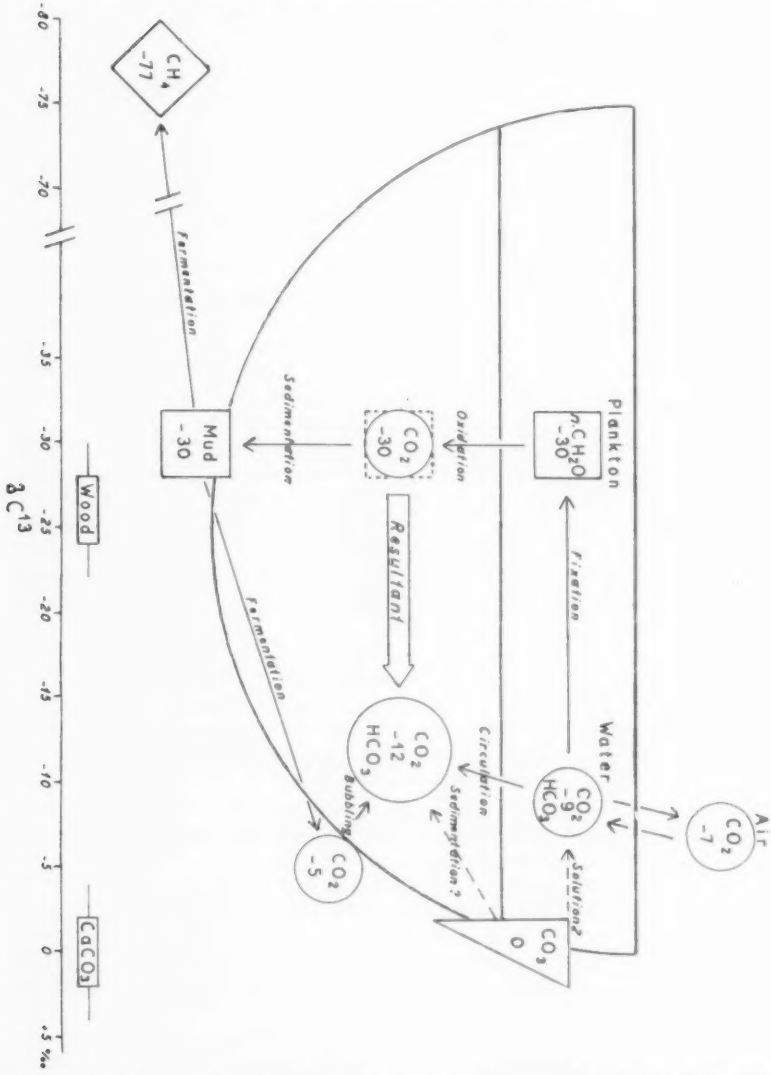

Fig. 1. Carbon cycle in a dimictic lake, based on C^{13} assays in Linsley Pond. The C^{13} contents of gases, deep-water CO_2, and other substances are plotted along the scale at the bottom, with reference to PDB-Chicago, a marine carbonate. Note break in scale at left.

metabolic or 30 percent of the total. Such values are distorted by contamination with carbonate carbon; and, as Linsley Pond appears to have a higher pelo-

metabolic contribution than Wononscopomuc's, and about the same as Queechy's, it is difficult to avoid the conclusion that Linsley, too, contains appreciable carbonate carbon.

Other considerations reinforce this conclusion, and lead to suspicion of the Quonnipaug results as well. In column 6 the estimated concentrations of pelometabolic CO_2 are referred to unit area of mud surface, on the assumption (fully justified for section *b* of the table) that the mean hypolimnetic CO_2 is distributed uniformly through the mean thickness of the hypolimnion. If it is then assumed that this heavy CO_2 has been produced by the mud during 100 days of stagnation, the estimated gas production is as shown in column 7. The smallest production is 10.3 ml/m².day, estimated for Quassapaug from the titration data.

Mud gases when collected in the water typically contain 1-2 percent CO_2; and a gas production of about $10 \div 0.02$ or 500 ml/m².day would not seem unreasonable; Conger (1943) measured 630 ml/m².day in a shallow lagoon, and from the data of Koyama (1954) we calculate at least 494 ml/m².day for Nakatsuna-ko and 335 ml/m².day for Kisaki-ko. The same arguments lead to grotesque results for the other lakes, however: it seems improbable that Quonnipaug produces between three and ten times as much gas as Quassapaug, and quite inconceivable that Linsley Pond should yield fifty to a hundred times as much. Relaxation of the secondary assumptions (length of stagnation, percentage of CO_2) does not remove the difficulty, and the basic assumption is plainly fallacious except for Quassapaug.

A note on radiocarbon

Carbonate carbon is as rich in C^{13} as is pelometabolic CO_2, but the two sources can easily be distinguished by the fact that carbonate rock is "infinitely" old and contains no C^{14}. Proof that Queechy Lake contains appreciable carbonate carbon was presnted by Deevey et al. (1954), and the same paper shows that Quassapaug, selected as a soft-water lake, contains little if any old carbon. A complete re-study is in progress, but some new data are mentioned here in order to substantiate the conclusions of the previous section. Three water samples have been analyzed for C^{14}; after normalization to a constant C^{13}/C^{12} ratio, following the notation of Broecker and Olson (1959) the results are:

		δC^{14}, ‰	δC^{13}, ‰	ΔC^{14}, ‰
Y-742	Linsley Pond, Oct. 20, 1958, 13.5 m	–81	–17.2	–100
Y-744	Queechy Lake, July 12, 1959, surface	–270	– 5.9	–311
Y-745	Queechy Lake, Sept. 27, 1959, 12 m	–282	–11.4	–316

These figures mean that Linsley's hypolimnion contains about 10 percent less radiocarbon, and Queechy's whole water column contains about 31 percent less, than do modern wood or carbonate. The modern atmosphere has gained about 20 percent C^{14} from nuclear tests, and lost about 3 percent through industrial pollution, but these processes are not likely to have changed the isotopic ratio of carbonate. We assume, pending direct tests, that the carbon fermented in mud is modern. It follows that, whereas all of the 30 percent or so of heavy CO_2 in Queechy could be derived from carbonate, only about a third of the heavy CO_2 in Linsley can be accounted for in this way. That is, if 30 percent

of the total in deep water is from a heavy source, and if 10 percent of the total is from carbonate, about 20 percent of the total, or about 40 percent of the metabolic CO_2, is pelometabolic. Such a figure could be reasonably compared with the estimate, between 4 and 20 percent of the metabolic CO_2, for Quassapaug, and as Linsley not only has a smaller and shallower hypolimnion, but is considered to be a more productive lake, the difference seems to be of the right order of magnitude.

The figure of 40 percent is smaller than the one computed by Hutchinson (1957) from oxygen and alkalinity data for the season of 1937, when the metabolic CO_2 anomaly represented about 50 percent of the metabolic CO_2. As Hutchinson pointed out, the inference that this amount is all pelometabolic fails to allow for any contribution from carbonate. The agreement between the estimates, made by widely different methods in different years, is astonishing, for Linsley Pond, like the other lakes, appears to have become more polluted since 1937. The typical summer oxygen curves suggest as much, and the metabolic CO_2 content of Linsley Pond as given in table 6 is more than twice as great as the mean figure of 21.8 mg/l, or 11.1 ml/l, calculated from Hutchinson's Table 87. Perhaps the pelometabolism, too, has been accelerated by cultural eutrophication, as Ohle's (1958) argument implies.

Autumnal circulation

Only two lakes have been studied after thermal stratification had broken down. One of these is Pocotopaug, East Hampton, Connecticut, (41°36′ N Lat, 72°30′ W Long), a very soft Eastern Highland lake for which no other C^{13} data are available. The lake is large (206.7 ha) and shallow (max. depth 11 m), and although it is stably stratified in summer the bottom temperature is high (ca. 14° C) and no proper hypolimnion is present. Autumnal circulation therefore begins earlier than in other lakes, and was well under way on October 27, 1959. Linsley Pond was also visited on November 27, 1959. The data are as follows:

Depth, m	Temp., °C	Total CO_2, ml/l	δC^{13}, ‰
	Pocotopaug, October 27, 1959		
0	13.7	3.30	−7.5
9.5	13.2	3.93	−6.6
	Linsley Pond, November 27, 1959		
0	8.0	31.3	−10.6
13.5	7.3	32.3	−10.6

Some heavy CO_2 may be present in the deeper water as a carryover from stagnation, but the dispersal of the remainder through the whole lake has not distorted the surface C^{13} beyond the value expected at any time of year. Although the epilimnion in general appears to be at equality (not in equilibrium) with the atmosphere, attention may be called to the occasional departures observed in Linsley Pond; values of δC^{13} as high as −4.9 per mil and as low as −11.2 per mil presumably record times of exceptionally intense photosynthesis and respiration, respectively, and departures as large as these are unlikely to result from admixture with hypolimnetic water having any reasonable δC^{13}.

Mud gases

Although we were obliged to encourage the formation of bubbles, it is a

spontaneous process at the pressures prevailing in these lakes. Ohle (1959), who was able to track methane bubbles with a fathometer, found them to appear rather suddenly when oxygen disappeared from the water, and considers them important in redistributing nutrients from a reduced hypolimnion. The usual method of collection, by jiggling, is nonquantitative for area sampled and for rate of production; moreover, it permits exchange between bubble and water, loss of CO_2 being especially likely. We resorted to this method because entrapment of cores (Koyama, 1954) yields too little gas, and experimental incubation (Allgeier, et al., 1932; Kusnezow, 1959) entails progressive isotopic fractionation.

The composition of such bubbles (table 7) gives some evidence of differential exchange: CO_2 is much less abundant than CH_4, and the low $Ar:N_2$ ratios (Oana, 1957) imply contamination[3] by N_2. The H_2 can hardly be a contaminant, however, and O_2 occurs only in traces except in spring. Since exchange can occur before as well as during collection, the relevant question is the degree to which it may alter the isotopic ratios. Despite some irregularities, the C^{13} data indicate an enrichment-depletion cycle of classic dimensions. The anaerobic production of CH_4 is coupled almost stoichiometrically to that of CO_2; with respect to the substrate (mud) with $\delta C^{13} = -30$ per mil, δC^{13} in the CH_4 approaches –80 per mil, while that in the CO_2 approaches zero. Some departures from this ideal or maximal fractionation, such as that of the CO_2 in Quassapaug, probably result from exchange with respiratory CO_2 in mud or water. The others may mean only that the CO_2 samples were too small for reliable assay; the highest values of δC^{13} tend to occur with the largest amounts of CO_2, and there appears to be no other relation with variation in composition, from lake to lake or with season or depth within lakes. All samples have doubtless lost some CO_2, but any fractionation during loss should work in the opposite direction, to enrich the C^{13} in the poorest samples.

Two recent studies confirm the extreme fractionation during methane fermentation. In unpublished data of Oana on the natural-gas field at Yamagata, Japan, δC^{13} in CH_4 averages –71.6 per mil, while that in CO_2 can be as high as +2.0 and averages –6.9 per mil. The weighted mean δC^{13} indicates a substrate, presumably petroleum, having a ratio of about –30 per mil, completing the analogy with lake mud. Rosenfeld and Silverman (1959) incubated a methane-producing culture derived from deep-sea mud, and when methanol was added the C^{13} in the CH_4 was depleted about 80 per mil, and that in the CO_2 was enriched at least 17 per mil, with respect to the methanol. The CH_4 of Linsley Pond in August, 1959, is the lightest natural carbonaceous substance yet analyzed, but the fractionation factor of 1.094 at 23°C, reported by Rosenfeld and Silverman, indicates that much lighter CH_4 is to be looked for.

All natural methane production is believed to be bacterial, and the reduced mud of lakes is one of the principal places where it occurs. The organisms responsible are various, and biochemically versatile; Stephenson (1949) gave an excellent review of their metabolism, and Kusnezow's work (1959) on lake

[3] Some fractionation of N-isotopes should result from nitrification, as some has been found by Ford and Hoering (1958) during N-fixation. The N^{15}/N^{14} ratio in mud gases would be interesting and might be helpful in distinguishing contaminated samples.

bacteriology is available in a German edition. Since we can specify neither the carbon source nor the identity of the bacteria, we cannot write all the reactions that are involved, but it is clear that those leading to methane and to CO_2 are coupled. The energy-source of methane bacteria is a reaction of the type

$$4\ H_2A + CO_2 = 4A + CH_4 + 2\ H_2O,$$

where A is any compound for which the organism possesses a dehydrogenase. Among known reactions of this type, at least two could be coupled with the formation of CO_2 and CH_4:

$$2\ C_2H_5OH + CO_2 = CH_4 + 2\ CH_3COOH;$$
$$CH_3COOH + CO_2 = CH_4 + 2\ CO_2;$$

in which ethyl alcohol is broken down to acetic acid, methane, and CO_2, and CO_2 is both a reactant and a product. These are not likely to be the main reactions proceeding in the mud of lakes, but the system is an analogous oxido-reduction. Isotopic studies with these bacteria show that their metabolism is more complex than the equations imply; for example, in the breakdown of ethyl alcohol, if C^{14}-labelled alcohol is supplied, the C^{14} appears in the cell substance and in the methane, but not in the acetic acid.

The important conclusion is that marked isotopic fractionation should result when the carbohydrates and hydrocarbons of lake mud are fermented anaerobically to CH_4 and CO_2. Even more important is the inference that this process is quantitatively important at the surface of an oxidized earth; it is customary to think of the methane of the primitive atmosphere as having all been oxidized, but an important part is evidently cyclic (Hutchinson, 1949, 1954). If 4 to 20 percent of the metabolic CO_2 in the hypolimnion of a lake as unproductive as Quassapaug is formed by fermentation, and if one or more (probably at least two) moles of CH_4 are formed at the same time for each mole of CO_2, the lakes and swamps of the world are producing methane on a colossal scale. And yet, in the process, the amount of fixed carbon that is reduced is negligible, and can make almost no contribution to diagnesis of sediments; at least in Koyama's lakes the daily loss of carbon as CO_2 and CH_4 amounts to only 7×10^{-4} percent (Nakatsuna) or 3.7×10^{-4} percent (Kisaki) of the organic carbon in the upper 40 cm of the sediments, or less than 0.3 percent per year. More carbon than this is annually lost from lakes when the benthic insects emerge (Vallentyne, 1952).

The discovery that fermentation CO_2 is isotopically enriched at the expense of methane raises an important question about groundwater. The carbonate-rich groundwater of the Heidelberg district, Germany, is unexpectedly rich in C^{14}, and the C^{13} in the water and in carbonate precipitated from it is depleted (about $-10‰$) with respect to carbonate rock (Münnich and Vogel, 1959; Vogel, 1959; Brinkmann et al., 1959). Antler bone from several archaeologic sites near Hamburg, Germany, has given anomalously young C^{14} ages, presumably for this reason (Münnich, 1957a, 1957b). Münnich (1958) and Vogel (1959) suppose that the contaminating carbon comes from modern humus, and becomes quantitatively more important than old carbonate owing to the very large pressures of CO_2 that can be built up at depth. Such an ex-

planation may be unnecessary if fermentation CO_2 is enriched in C^{14} as it is in C^{13}.

Our analyses of cave limestones and groundwater will be published elsewhere; in general, we can confirm the depletion of secondary $CaCO_3$ in C^{13}, but in two analyses of Saratoga Springs waters, where large amounts of CO_2 are involved, we find enrichment instead, δC^{13} averaging −2.3 per mil for the CO_2, and + 4.3 per mil for the carbonate precipitate. Such a result is also predicted by the Münnich-Vogel model, however, and the question must remain open until the C^{14} content of such CO_2 can be measured directly.

Pondweeds and molluscs

The δC^{13} values for plants and animals in the five lakes, given in tables 1 through 5, contain no surprises, if one accepts the evidence that aquatic plants can utilize bicarbonate directly as a carbon source. As Craig's (1953) data also show, the organic carbon of submerged aquatics (*Myriophyllum*, *Nitella*, *Potamogeton*, *Chara*) is enriched in C^{13} in proportion to the hardness of the water. The lowest values are thus found in Quassapaug, the highest in Wononscopomuc and Queechy. On the other hand, such a floating aquatic as *Nuphar* (yellow water lily) appears to take most of its carbon from the air, and its δC^{13} is about that of a terrestrial plant and thus lower than that of the submerged aquatics, except in Quassapaug, where it is about the same. The same difference was already apparent in the C^{14} data from Queechy and Quassapaug (Deevey et al., 1954), in that the submerged aquatics incorporated (old) carbonate carbon in Queechy, while the emergent aquatics did not. Because some atmospheric air is involved in any case, however, the effect of photosynthesis is to make the organic carbon of an aquatic plant richer in C^{14} than the bicarbonate of a hard water, while it is poorer than the bicarbonate in C^{13}.

This latter, second-order effect is also apparent in the C^{13} content of the shells, which are slightly depleted with respect to the water, more definitely so in the two hardwater lakes. (No molluscs were found in Quassapaug). It is generally true, though difficult to explain on physiologic grounds, that shell displays no fractionation of carbon isotopes with respect to the medium, and the data of Broecker and Olson (1959) show that this is as true of freshwater as of marine shells. The slight depletion that we observe is probably to be attributed to depletion of C^{13}, presumably by respiratory CO_2, in the local medium (shallow-water mud) where the molluscs live. No doubt for the same reason, the shell of Queechy molluscs was found in 1954 to be richer in C^{14} than surface water. The absence of fractionation, apart from this small effect, is striking when one realizes that the organic matter of molluscs, not only the flesh but the horny periostracum of the shell itself (table 3) is as depleted as any other organic matter that has undergone a metabolic cycle.

MEROMICTIC LAKES

We had anticipated marked fractionation of C^{13} in the monimolimnion, or permanently unmixed lower portion of a meromictic lake. Through the kindness of Gerald Lauer and W. T. Edmondson, of the University of Washington, we were able to measure samples from two lakes of this type. The meromixis arises

from quite different causes, as discussed by Hutchinson (1957). In Soap Lake, a highly saline lake (described by Anderson, 1958) in the Grand Coulee of eastern Washington, a recent rise of water table has brought fresher water (salinity ca. 35 per mil) to overlie the older, denser water mass, the salinity of which is 144 per mil. In Hall Lake, near Seattle, the solutes that make the bottom layer denser are the ordinary substances that accumulate in lakes during stagnation, and the meromixis probably began during an exceptional spring when circulation was incomplete. The ages of the bottom layers are unknown, but probably do not exceed a few years in Hall Lake or a few decades in Soap Lake.

The data are as follows:

Soap Lake, June 29, 1959		
Depth, m	0	25
Total CO_2, ml/l	2300	21,050
δC^{13}, ‰	+0.1	−2.4
Hall Lake, July 1, 1959		
Depth, m	0	15
Total CO_2, ml/l	19.3	35.6
δC^{13}, ‰	−9.8	−12.6

There is little or no fractionation, but the reasons are obvious on reflection. The monimolimnion is devoid of oxygen; as fermentation proceeds, heavy CO_2 accumulates, in Hall Lake at any rate, until the isotopic ratio of the total CO_2 approaches that of the surface water. Thus Hall Lake is a somewhat extreme example of the type of Linsley Pond, and its soft surface waters are nearly as depleted in C^{13} by admixture of respiratory CO_2 as is the monimolimnion after several years of stagnation. Soap Lake, on the other hand, is an excessively alkaline arid-region lake of the kind that also occurs in the Lahontan basin, and contains relatively huge amounts of sodium carbonate and bicarbonate; its surface C^{13} content is therefore that of carbonate rock, and the δC^{13} of the older water mass, before it became a monimolimnion, was probably also close to zero. For comparison, we note that direct examination of C^{13} content has been made on the water of two lakes in the Lahontan basin (Broecker and Walton, 1959; Broecker and Olson, 1959): Pyramid Lake gave −5.9 per mil, and Mono Lake +4.3 per mil. Even if CO_2 formed by fermentation in such an environment has a positive δC^{13}, which we do not know, its proportion can hardly be expected to influence the several thousand ml/l of carbonate CO_2 inherited from the older lake.

SEARLES LAKE, CALIFORNIA

The deposits of pluvial Searles Lake, the Parting Mud layer, lie interbedded with thick deposits of evaporites, dominated by trona. The recent history of this lake has been described by Flint and Gale (1958), who reported radiocarbon dates from several laboratories. At least twice in the last 46,000 years or more the basin was occupied by a deep, relatively freshwater lake, recording a pluvial time. From about 30,000 to about 24,000 B.P., and again from about 10,000 B.P. to the present, decidedly nonpluvial conditions prevailed and many feet of evaporites, the Lower and the Upper Salt beds, were formed as the lake evaporated to dryness.

Yale radiocarbon dates at four levels within the Parting Mud have been published by Deevey et al. (1959) as well as by Flint and Gale. We give here the C^{13} analyses, including some on the Lower and Upper Salt beds for which C^{14} analyses have not been obtained:

		Carbonate	Organic
Y-351	Upper Salt. Well X-10, 37.7 to 38.7 ft	+2.4‰	
Y-352	Upper Salt. Well K-10, 55.7 to 56.6 ft	+1.4	
Y-353	Upper Salt. Well X-10, 68.8 to 69.7 ft	+1.1	
Y-574	Parting Mud, top 8 in. Well X-20, 78 ft 10 in. to 79 ft 6 in.	+1.9	–26.7‰
Y-575	Parting Mud, one-third down. Well X-20, 81 ft 8.5 in. to 82 ft 4 in.	+0.2	–23.4
Y-576	Parting Mud, two-thirds down. Well X-20, 85 ft 4 in. to 86 ft 2 in.	+2.2	–19.2
Y-577	Parting Mud, bottom. Well X-20, 89 ft 11 in. to 90 ft 7 in.	+0.8	–14.1
Y-348	Lower Salt. Well X-16 98.0 to 98.9 ft	+2.5	
Y-349	Lower Salt. Well X-14 102.6 to 102.9 ft	+3.0	
Y-350	Lower Salt. Well X-14 105.6 to 106.7 ft	+1.1	

The C^{13} contents of the carbonate fractions, which are the only analyzable fractions of the trona beds, present nothing remarkable. These are almost the only available analyses of a sodium carbonate mineral, and it is hard to say whether the slight positive anomaly is characteristic of such material; recent tufa ($CaCO_3$) from the shore of Pyramid Lake (Broecker and Olson, 1959) is even more markedly enriched, with δC^{13} of +6.3 per mil, and some marine limestones are also consistently positive. The values for the organic fractions of the Parting Mud, however, are extraordinary in showing systematic depletion with time. We can only suppose that as the lake re-formed, early in the last pluvial time, its waters redissolved large quantities of the pre-existing carbonate, and quickly came to have a high C^{13} content. As the lake became more normal, i.e. as its bottom came to be covered with silt and organic mud instead of trona and halite, the usual processes of photosynthesis and respiration produced a mud that was normally depleted in C^{13} with respect to carbonate and air. It is remarkable, if this explanation is correct, that the total thickness of Parting Mud, at least as recorded in borings near the center of the basin, is only 10 to 13 feet.

If this mechanism is the usual one when pluvial lakes re-form and dissolve their old salt deposits, it should seriously distort both the C^{13} and the C^{14} ratios, perhaps more especially the latter. We therefore find it difficult to accept *in toto* the brilliant arguments of Broecker and Walton (1959), who explicitly discounted the redissolution of older carbonate in using the observed C^{14} ratios to calculate the invasion rate of atmospheric CO_2 into the waters of the Lahontan and Bonneville basins. It must be admitted, however, that the radiocarbon dates of organic and inorganic fractions of the Parting Mud are not significantly different (Deevey et al., 1959). If there were no appreciable time lag between the last evaporation and the formation of a new lake, the redissolved carbonate would not be old enough to affect the dating.

PRECAMBRIAN BANDED IRON ORES

Our attention was drawn to the iron-banded cherts of Middle Huronian age by the suggestion of Hough (1958) that these famous rhythmites are pri-

mary sediments, laid down in a warm monomictic lake. The iron, which occurs in all states of oxidation from FeS to Fe_2O_3, is supposed by Hough to have been mobilized in ferrous form in a reduced hypolimnion at a moderately low pH, and to have been deposited mainly as carbonate (siderite) and oxides (limnonite, magnetite, and hematite), according to the redox potential prevailing during annual circulation periods. As Hough pointed out, reducing conditions are a normal feature of hypolimnia, and such deposits, if they are lacustrine (and if they are sediments) cannot be used as evidence that an oxygen-rich atmosphere had not yet evolved; Geijer (1956) gave evidence that the fully oxidized hematite in similar deposits in Sweden is primary. On the other hand, since organic matter is the most probable reducing agent in such an environment, the iron compounds are presumably biochemical fossils of some sort.

These considerations would also apply if the environment of deposition were marine, as most geologists have assumed (James, 1954). A hypersaline lagoon could provide the required anoxia, though the near-absence of Ca, Mg, and Na rather favors the idea of freshwater origin. The C^{13} assay could not be expected to distinguish between marine and freshwater alternatives, but we thought it possible that fractionation between the iron- (and carbonate-) rich and the silica-rich layers might confirm or deny Hough's hypothesis of seasonal lamination. Through the kindness of Harold L. James, U. S. Geological Survey, we have been able to analyze four specimens (a similar specimen provided by J. L. Hough contains too little carbon for an accurate analysis). The data are as follows:

		δC^{13}, ‰
1.	Chert layer, chert-siderite; Iron River district, Michigan	–10.7
2.	Siderite layer; Iron River district, drill-hole 207	–13.7
2a.	Free carbon from same layer as 2	–30.5
3.	Chert layer, chert-siderite; Athens mine, Marquette district, Michigan	–9.4
4.	Speckled siderite layer, bedded iron fm; Tyler Fork, Gogebic district, Michigan; (carbonate probably entirely diagenetic)	–14.8

The samples are relatively large, and must entail some stratigraphic blurring; their main use is to show that the problem is a real one, deserving much more careful investigation. All are markedly depleted in C^{13}, not only in respect to marine and other carbonates deposited in oxidizing environments (including the Grenville and other Precambrian limestones; Craig, 1953), but in respect to mean "crustal" carbon, which Wickman (1956) assigned a value corresponding to –5.7 per mil on our scale. Supposedly primary siderites contribute importantly to Wickman's mean, together with diamonds and selected graphites, and there is little doubt that the Michigan siderites are anomalously low. Moreover, although the difference is not statistically significant, the iron-rich layers are depleted with respect to the silica-rich layers. Both kinds of anomaly, the former more certainly than the latter, give evidence in favor of Hough's interpretation. The free carbon in specimen 2 has the ratio expected of organic matter, but it is not necessary to debate the complex question of the C^{13} ratio of graphite to see in the other data a strong indication that organic or fixed carbon was present in Middle Huronian time, very possibly in the oldest known stratified lake(s).

CONCLUSION

The studies reported here raise questions concerning the metabolism of lakes, including the gaseous metabolism of the mud, which clearly demands continuing and more exact study in the field and in the laboratory. They are also of more than incidental interest in radiocarbon dating, not only of such materials as gyttja and freshwater shell, which may fossilize with more or less C^{14} than their C^{13} assays suggest, but of organic materials formed in closed basins containing extensive carbonate deposits, and of carbonates formed from groundwater that may have incorporated fermentative CO_2.

We had hoped to turn up something of even greater interest: an isotopic-assay method for distinguishing seasonally laminated lacustrine deposits from other rhythmites. In this we were disappointed; the problem is much more complex than we suspected when isotopic fractionation was first demonstrated in the hypolimnion of Linsley Pond. Some fractionation is evidently to be expected in a normal hypolimnion, and if alterations of C^{13} ratio are found between successive laminae in a single deposit, it may be legitimate to infer that oxidation and reduction have been partly segregated according to season, probably by thermal stratification. But the converse is probably not true: fermentation and methane-bubbling tends to obliterate the C^{13} fractionation, and the more productive the lake, or the more it tends toward meromixis for whatever reason, the less likely are we to see the effect of seasonal stagnation in the C^{13} contents of sediments. The presence in the basin of large amounts of carbonate, whether old or equilibrated with air, will inevitably lead to further complications. Organic and carbonate-rich laminae that are known on other grounds to be seasonal will surely repay investigation, and we intend to pursue the matter. We suspect, however, that seasonally diagnostic fossils will continue to furnish the best evidence of seasonal lamination in rhythmites.

TABLE 1

Quassapaug. Limnological and C^{13} data

Location: Middlebury, Conn. 41° 32′ N Lat, 73° 09′ W Long
Geologic province: Western Highland (Paleozoic crystallines)
Area: 117.2 ha
Mean depth: 8.7 m
Max. depth: 20 m
Volume ratio, hypolimnion to epilimnion: 0.35
Area to volume ratio, hypolimnion: 0.219 m^{-1}

	Depth, m	Temp., °C	Total CO_2, ml/l	δC^{13}, ‰
Sept. 19, 1958	0	19.7	1.24	—
	18	7.0	5.61	−22.6
Oct. 5, 1958	0	16.8	1.74	−8.7
	16	7.2	5.82	−21.9
	19	7.0	6.41	−20.8
June 15, 1959	0	19.7	1.84	—
	19	7.8	8.22	−17.5
Sept. 24, 1959	0	21.8	3.18	—
	17.5	7.8	14.4	−22.4
Net plankton, Sept. 1959				−26.4
Myriophyllum, Oct. 1958				−27.5
Nitella, Oct. 1958				−26.4
Nuphar, Oct. 1958				−23.7

Organic mud, 20 m, 1937	−29.2
Organic mud, 19 m, 1959, top 10 cm	−28.3
Organic mud, 19 m, 1959, 10-20 cm	−27.9

Mean total CO_2 by titration, ml/l:

	Epilimnion	Hypolimnion
June 15, 1959	2.3	5.8
Sept. 24, 1959	2.0	7.7

TABLE 2

Quonnipaug. Limnological and C^{13} data

Location: Guilford, Conn. 41° 23.5′ N Lat, 72° 42′ W Long
Geologic province: contact, Eastern Highland-Triassic Lowland
Area: 46.6 ha
Mean depth: 5.1 m
Max. depth: 14.8 m
Volume ratio, hypolimnion to epilimnion: 0.72
Area to volume ratio, hypolimnion: 0.191 m^{-1}

	Depth, m	Temp., °C	Total CO_2, ml/l	δC^{13}, ‰
Oct. 12, 1958	0	15.1	6.12	−7.9
	13	8.5	8.66	−17.4
	14	8.4	13.2	−17.1
June 22, 1959	0	23.7	12.2	−7.5
	13.5	8.2	20.8	−17.3
Sept. 15, 1959	0	21.4	18.9	−11.6
	0		13.8*	−8.1*
	13.5	9.0	31.1	−16.9
Potamogeton, Oct. 1958				−14.4
Nuphar, Oct. 1958				−25.0
Elliptio, Oct. 1958, shell				−9.0
flesh				−28.5
Organic mud, 1959, 14 m				−28.5

Mean total CO_2 by titration, ml/l

	Epilimnion	Hypolimnion
June 22, 1959	9.1	14.6
Sept. 15, 1959	11.2	19.3

* duplicate rejected on suspicion of incomplete extraction

TABLE 3

Linsley Pond. Limnological and C^{13} data

Location: Branford, Conn. 41° 19′ N Lat, 72° 47′ W Long
Geologic province: Triassic Lowland
Area: 9.4 ha
Mean depth: 6.7 m
Max. depth: 14.8 m
Volume ratio, hypolimnion to epilimnion: 0.67
Area to volume ratio, hypolimnion: 0.248 m^{-1}

	Depth, m	Temp., °C	Total CO_2, ml/l	δC^{13}, ‰
Sept. 19, 1958	0	20.0	16.7	−8.5
	14	6.9	23.0	−13.9
Oct. 20, 1958	0	14.1	13.9	−11.2
	13.5	6.9	26.8	−17.2
April 27, 1959	0	—	36.2	−7.8
May 17, 1959	0	17.8	12.4	−8.1
	14	7.1	34.0	−18.0*

June 4, 1959	0	23.8	23.3	–8.8
	14	7.0	38.9	–15.2
Aug. 24, 1959	0	25.2	28.6	–8.8
	13.5	7.4	48.9	–16.1
Oct. 2, 1959	0	22.0	28.4	–4.9
	13.5	7.4	51.9	–12.7
Oct. 21, 1959	0	14.9	30.0	–10.6
	9	8.8	48.7	–12.8
	11	8.0	47.6	–12.7
	12	7.8	52.7	–12.4
	13.5	7.7	47.2	–12.1
Oct. 29, 1959	0	13.0	30.1	–10.3
	13	7.9	54.2	–11.9
Net plankton, Oct. 2, 1959				–28.0
Potamogeton, Sept. 1958				–20.6
Nuphar, Sept. 1958				–22.9
Elliptio, no. 1, Oct. 1958, shell				–11.2
flesh				–31.8
no. 2, Oct. 1958, shell				–12.2
flesh				–30.6
no. 3, Oct. 1958, periostracum				–28.9
Organic mud, 1958, 14 m				–30.9

Mean total CO_2 by titration, ml/l

	Epilimnion	Hypolimnion
June 4, 1959	14.5	22.2
Aug. 24, 1959	23.5	29.2
Oct. 2, 1959	24.4	35.2

* duplicates dissimilar, –22.2 and –15.8‰

TABLE 4

Wononscopomuc Lake. Limnological and C^{13} data

Location: Salisbury, Conn. 41° 57.5′ N Lat, 73° 27′ W Long
Geologic province: Stockbridge marble
Area: 142.8 ha
Mean depth: 11.1 m
Max. depth: 31 m
Volume ratio, hypolimnion to epilimnion: 0.63
Area to volume ratio, hypolimnion: 0.119 m^{-1}

	Depth, m	Temp., °C	Total CO_2, ml/l	δC^{13}, ‰
Sept. 29, 1958	0	18.1	23.6	–6.1
	20	5.9	27.8	—
	23	5.5	31.1	–11.5
July 22, 1959	0	26.8	47.1	–4.7
	28	5.3	63.6	–10.1
Sept. 18, 1959	0	19.8	47.1	–4.0
	29	5.7	67.5	–10.3
Net plankton, Sept. 1959				–32.2
Potamogeton, Sept. 1958				–13.7
Vivipara, July 1959, shell				–8.8
Organic mud, July 1959, 29 m, top 10 cm				–30.3
10-20 cm				–30.4

Mean total CO_2 by titration, ml/l:

	Epilimnion	Hypolimnion
July 22, 1959	24.1	27.6
Sept. 18, 1959	32.4	39.6

TABLE 5

Queechy Lake. Limnological and C^{13} data

Location: Canaan, N. Y. 42° 24.5′ N Lat, 73° 25′ W Long
Geologic province: Stockbridge marble
Area: 60.8 ha
Mean depth: 6.9 m
Max. depth: 13.8 m
Volume ratio, hypolimnion to epilimnion: 0.37
Area to volume ratio, hypolimnion: 0.276 m^{-1}

	Depth, m	Temp., °C	Total CO_2, ml/l	δC^{13}, ‰
Sept. 28, 1958	0	18.0	19.3	−9.1
	11	8.9	30.2	−12.9
	12	7.5	43.9	−13.1
July 12, 1959	0	24.9	42.0	−5.9
	12.2	6.3	81.6	−11.7
Sept. 27, 1959	0	19.5	42.8	−5.1
	12	7.0	82.8	−11.4
Net plankton, Sept. 1959				−35.5
Chara, Sept. 1958				−15.6
Potamogeton, Sept. 1958				−14.9
Elliptio, Sept. 1958, shell				−10.8
flesh				−29.8
Organic mud, 1937, 13 m				−32.5
Organic mud, 1938, 7 m, organic fraction				−27.5
carbonate fraction				−8.5

Mean total CO_2 by titration, ml/l:

	Epilimnion	Hypolimnion
July 12, 1959	22.0	38.7
Sept. 27, 1959	26.9	46.5

TABLE 6

Fermentation CO_2 in the hypolimnion in late summer 1959, estimated from the proportion of C^{13}-rich CO_2

	1 Residual (surface) CO_2, ml/l	2 Metabolic (deep-water) CO_2, ml/l	3 Heavy (pelo-metabolic?) CO_2, ml/l	4 (3) as % of (2)	5 (3) as % of total deep CO_2	6 (3) per unit area, ml/m^2	7 (3,6) as produced in 100 days, ml/m^2. day
a. From gasometric data on bottom water							
Quassapaug	1.8	12.5	2.57	20.5	17.9	11700	117
Quonnipaug	12.2	18.9	5.48	29.0	17.6	28800	288
Linsley	23.3	28.6	15.6	54.5	30.0	63000	630
Wononscopomuc	47.1	20.4	12.3	60.2	18.2	102000	1020
Queechy	42.0	40.8	25.2	61.7	30.8	91300	913
b. From titrable CO_2 in whole hypolimnion							
Quassapaug	2.3	5.4	0.225	4.2	2.9	1030	10.3
Quonnipaug	9.1	10.2	2.05	20.1	10.6	10800	108
Linsley	14.5	20.7	11.8	56.8	33.4	47700	477
Wononscopomuc	24.1	15.5	10.3	66.5	26.0	86500	865
Queechy	22.0	24.5	15.5	63.3	33.3	56100	561

TABLE 7

Mud gases. Percentage composition and δC^{13}

	N_2 %	CH_4 %	CO_2 %	Ar %	H_2 %	O_2 %	δC^{13}, CH_4 ‰	δC^{13}, CO_2 ‰
Quassapaug								
5. X .58, shallow	89.5	7.7	2.0	0.5	0.3	0.0	–62.2	
24.IX.59, shallow	52.7	43.9	2.16	0.93	0.25	0.09	–68.0	–25.2
Quonnipaug								
15.IX.59, shallow	66.3	30.7	1.60	1.11	0.24	0.05	–58.9	
Linsley								
27.IV.59, shallow	91.4	2.78	0.58	0.87	0.00	4.33	–65.6	
17.V.59, shallow	93.8	3.49	0.33	0.97	0.00	1.38	–73.2	
17.V.59, deep	32.1	65.2	0.38	0.47	0.20	1.65	–77.4	–20.6
4.VI.59, deep	26.7	71.5	1.01	0.48	0.27	0.07	–77.8	–7.1
24.VIII.59, deep	23.5	74.8	0.90	0.41	0.31	0.08	–80.2	–6.1
8.X.59, deep	26.4	71.5	1.17	0.50	0.26	0.05	–77.5	–7.2
Wononscopomuc								
22.VII.59, shallow	16.0	77.7	5.54	0.38	0.32	0.06	–59.7	–6.1
18.IX.59, shallow	22.6	74.1	2.4	0.50	0.38	0.02	–60.8	–14.7
Queechy								
11.VII.59, shallow	57.8	35.2	4.36	0.91	0.38	0.07		–4.2
27.IX.59, deep	35.5	61.7	1.77	0.71	0.31	0.01	–57.2	–15.2

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THE NATURE OF RESIDUAL LIQUIDS FROM CRYSTALLIZATION, WITH DATA ON THE SYSTEM NEPHELINE-DIOPSIDE-SILICA

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ABSTRACT. The phase-equilibrium relations in the system nepheline-diopside-silica have been determined. The join albite-diopside, although it is not binary, divides this system into two parts, a silica-rich portion with silica as one of the solid phases in completely crystallized mixtures and a nepheline-rich portion with nepheline as one of the solid phases in the final crystallization product. In the system nepheline-diopside-silica the primary phases are diopside (probably an aluminous diopside solid solution), complex nepheline solid solutions, carnegieite solid solutions, sodic plagioclases, forsterite, tridymite, and cristobalite. The join jadeite-diopside is not unique in this system at 1 atmosphere pressure and jadeite is not encountered as a solid phase.

The results support Bowen's contention that through fractionation a magma is impoverished in the early-crystallizing minerals (in this case diopside) and yields residual liquids rich in the alkali aluminosilicates.

The nature of the albite-diopside join indicates that no single anhydrous composition can yield a liquid which through fractionation produces both a silica-rich residue and a nepheline-rich residue.

There is a reaction relation between olivine and diopside. This reaction relation should be as important in the alkali basalts as the olivine-hypersthene reaction relation is in the tholeiitic basalts.

RESIDUAL LIQUIDS FROM CRYSTALLIZATION

Some years ago Bowen (1937) discussed the nature of residual liquids from crystallization of complex silicate systems and called the system $NaAlSiO_4$–$KAlSiO_4$–SiO_2, petrogeny's "residua" system. By the investigation of a number of relatively simple systems combining one of the early-crystallizing minerals of rocks with late-crystallizing alkali alumino silicates, we may reach some general conclusions regarding the crystallization of complex mixtures. The relations of these simple systems to petrogeny's "residua" system are shown in figure 1.

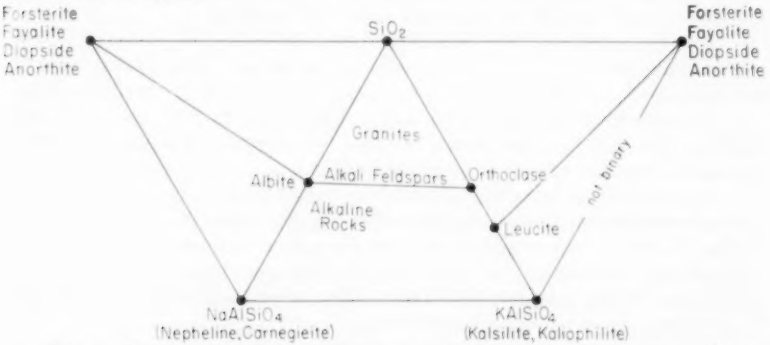

Fig. 1. Diagram showing interrelation of systems involving both early- and late-crystallizing minerals. The system $NaAlSiO_4$–$KAlSiO_4$–SiO_2 is petrogeny's "residua" system.

Petrographic evidence indicates that olivine is one of the earliest of the

common rock-forming minerals to crystallize from rocks. Olivines are for the most part solid solutions of forsterite ($2MgO \cdot SiO_2$ or Mg_2SiO_4) and fayalite ($2FeO \cdot SiO_2$ or Fe_2SiO_4). A pyroxene may be one of the next minerals to crystallize. Diopside ($CaO \cdot MgO \cdot 2SiO_2$ or $CaSiO_3 \cdot MgSiO_3$) is a simple pyroxene. One of the next minerals to crystallize is calcic plagioclase, a high-lime feldspar. Anorthite ($CaO \cdot Al_2O_3 \cdot 2SiO_2$ or $CaAl_2Si_2O_6$) is the high-lime end member of the plagioclase series, anorthite ($CaAl_2Si_2O_6$)—albite ($NaAlSi_3O_8$). By combining the early-crystallizing minerals olivine, pyroxene, and lime feldspar with potash aluminosilicates and also with soda aluminosilicates, we get a series of simple systems, which have been investigated. All these systems yield residual liquids from crystallization which are very rich in alkali aluminosilicates and lead us to the conclusion that residual liquids from fractional crystallization in complex systems approach the low-temperature trough[1] in petrogeny's "residua" system $NaAlSiO_4$–$KAlSiO_4$–SiO_2. A phase-equilibrium diagram for this system has been given by Schairer (1950).

The status of the simple systems shown in figure 1 is as follows:

Leucite-forsterite-silica.—Schairer (1954) studied this system and a series of joins in the quaternary system K_2O–MgO–Al_2O_3–SiO_2. The two volumes leucite-corundum-spinel-silica and leucite-forsterite-spinel-silica constitute a large portion of this quaternary system and include the portion of most interest to the petrologist who is concerned with the origin of igneous rocks or with the mineral assemblages of metamorphic rocks. All the crystalline phases which are found in these two volumes are common rock-forming minerals or accessory minerals of either igneous or metamorphic rocks. The goal toward which the composition of the liquid phase proceeds during crystallization is essentially the same in the system leucite-forsterite-silica and in the two large volumes. This goal is a mixture of potash feldspar and silica, in these soda-free compositions, with a small amount of a magnesian mineral.

Leucite-fayalite-silica.—This system was studied by Roedder (1951) at the University of Utah. The goal of crystallization is also the same mixture of potash feldspar and silica with a small amount of fayalite.

Leucite-diopside-silica.—This system was studied by Schairer and Bowen (1938). The goal of crystallization is also this same mixture of potash feldspar and silica with a small amount of a magnesian mineral, in this case diopside.

Leucite-anorthite-silica.—This system was studied by Schairer and Bowen (1947). The goal of crystallization is again this same mixture of potash feldspar and silica, with a small amount of anorthite.

Nepheline-forsterite-silica.—Greig has studied this system at the Geophysical Laboratory, but the results are not ready for publication. Schairer and Yoder are studying a series of joins in the quaternary system Na_2O–MgO–Al_2O_3–SiO_2. Preliminary results on some of these joins have already appeared (Schairer, 1957a; Schairer and Yoder, 1958). Residual liquids from crystallization are rich in soda aluminosilicates.

[1] The trough of petrogeny's "residua" system was illustrated by Bowen as crossing the join albite-orthoclase. From the work of Chayes (1956) and Tilley (1957) on natural rocks which may be represented for the most part in the silica-rich and nepheline-rich portions of the "residua" system, it appears that the natural rocks group along two troughs approximately parallel to the albite-orthoclase join.

Nepheline-fayalite-silica.—This system was studied by Bowen and Schairer (1938). Residual liquids from crystallization are rich in sodium aluminosilicates and the join fayalite-albite divides the system into two parts. In one of these silica is one of the solid phases and in the other, nepheline.

Nepheline-diopside-silica.—The phase-equilibrium data for this system are presented here in this paper.

Nepheline-anorthite-silica.—This system has been studied by Schairer (1957b), who gave a preliminary phase-equilibrium diagram. The data are complete but the results have not been written up for publication. Residual liquids in the portion nepheline-albite-anorthite are rich in sodium aluminosilicates and are nepheline-bearing. The crystallization goal in the portion albite-anorthite-silica is a mixture of soda feldspar and silica in these potash-free compositions.

Special attention is called to the fact that in the systems nepheline-forsterite-silica, nepheline-fayalite-silica, and nepheline-anorthite-silica the joins albite with forsterite, fayalite and anorthite, respectively, divide these systems into two parts, a siliceous part with silica as one of the solid phases and an alkalic part with nepheline as one of the solid phases. On the other hand, in the systems leucite-forsterite-silica, leucite-fayalite-silica, leucite-diopside-silica and leucite-anorthite-silica, because of the incongruent nature of the melting of leucite, the joins potash feldspar with forsterite, fayalite, diopside, and anorthite, respectively, do not in a similar manner divide these systems.

THE SYSTEM NEPHELINE-DIOPSIDE-SILICA

Experimental methods.—In the preparation of silicate mixtures containing Na_2O, special care must be used to avoid loss of Na_2O and consequent departure from composition. Pure dry crystalline sodium disilicate was prepared from Na_2CO_3 and purified quartz by methods already described by Schairer and Bowen (1956). The mixtures in nepheline-diopside-silica were prepared by fusing together in platinum crucibles appropriate weighed quantities of sodium disilicate, pure Al_2O_3 (T61 free from Fe_2O_3 from the Aluminum Company of America, ignited 3 hours at 1300°C), pure crystalline synthetic diopside, and purified quartz. It was necessary to fuse a mixture several times (usually five to eight) with intermediate grinding to secure a homogeneous glass. The homogeneity of the glass was checked under the microscope by the uniform index of refraction of the glass grains. The mixtures were crystallized completely at appropriate temperatures before proceeding with the quenching experiments. Some of the more siliceous compositions required several months before crystallization was complete. The crystalline phases encountered in this system were identified by their optical properties under the petrographic microscope or by their X-ray powder patterns or by both methods. Equilibrium was reached in a few hours in this system at temperatures above 1200°C. Below this temperature runs up to two weeks' duration were necessary. The temperature was measured with a Pt-Pt 90 Rh 10 thermoelement calibrated frequently at several fixed points defined in degrees Centigrade as follows: NaCl melting point, 800.4°; gold melting point, 1062.6°; diopside melting point, 1391.5°; and $CaSiO_3$ melting point, 1544°.

Limiting systems.—Data for the system diopside-silica are given in two previous papers. Bowen (1914) showed the eutectic relations between diopside and tridymite at 84 weight percent diopside at the temperature 1362°C. Greig (1927) gave data for the region of two immiscible liquids in the ternary system CaO–MgO–SiO_2, from which the binary relations for diopside–SiO_2 may be interpolated.

Data for the system $NaAlSiO_4$–SiO_2 are given in two previous papers. Greig and Barth (1938) determined the relations in the portion nepheline-albite. Schairer and Bowen (1956) determined the relations in the portion albite-silica.

The system nepheline-diopside was studied by Bowen (1922). Recently Schairer, Yagi, and Yoder made a more detailed examination of this system and a manuscript on their results is in preparation. Their liquidus results are incorporated here in the phase diagram for nepheline-diopside-silica with their kind permission. The system nepheline-diopside is not binary. Forsterite appears on the liquidus surface and melilite appears as a solid phase at subliquidus temperatures.

The join albite-diopside was studied in part by Bowen (1915), who did not prepare and study mixtures richer in albite than 90 percent. He assumed that the system was binary with a eutectic at about 97 weight percent albite at about 1085°C. As we shall see later, this system is not binary and sodic plagioclase appears on the liquidus surface.

Phase-equilibrium diagram for nepheline-diopside-silica.—The critical results of quenching experiments on fifty-seven synthetic mixtures are recorded in table 1 and the phase-equilibrium diagram[2] is given here as figure 2. All

[2] Preliminary diagrams for this system and for the join jadeite-diopside were given by Yoder (1950).

TABLE 1

Results of quenching experiments

(a) Mixtures except those on the lines diopside with jadeite or with albite

Composition			Results of microscopic and/or x-ray examination		
Nepheline	Diopside	Silica	Primary phase	Liquidus temp.,°C	Other data
15	80	5	diop	1328	
25	65	10	diop	1288	
40	55	5	diop	1228	ness in 1152°; B.M. 970 ± 10°
15	55	30	diop	1313	tridy in 1272°
10	55	35	tridy	1422	diop in 1323°
55	40	5	fo	1200	diop in 1163°; ness in 1153°; fo out 1132°; B.M. 975 ± 10°
20	40	40	diop	1283	tridy in 1272°
10	40	50	cristo	1527	tridy in 1470 ± 10°; diop in 1312°
60	35	5	fo	1183	diop in 1168°; ness in 1163°; fo out 1157°; B.M. 1045 ± 5°
55	35	10	fo	1185	diop in 1178°; ness in 1138°; fo out 1133°; B.M. 965 ± 10°
65	30	5	ness	1198	fo in 1183°; diop in 1168°; fo out 1158°; B.M. 1090 ± 5°

60	30	10	fo	1160	diop in 1153°; ness in and fo out 1138°; B.M. 970 ± 10°
55	30	15	diop	1175	ness in 1123°; B.M. 980 ± 10°
45	30	25	diop	1218	plag in 1123°; ness in 1038°; B.M. 970 ± 10°
20	30	50	tridy	1298	diop in 1278°
65	25	10	ness	1180	fo in 1152°; diop in 1138°; fo out 1133°; B.M. 1020 ± 5°
30	25	45	diop	1248	plag in 1078°; cristo* in 1073°; B.M. 1020 ± 10°
75	20	5	ness	1248	fo in 1178°; diop in 1158°; fo out 1128°; B.M. 1030 ± 5°
65	20	15	ness	1143	diop in 1128°; B.M. 1025 ± 5°
60	20	20	diop	1140	ness in 1108°; plag in 1070°; B.M. 985 ± 5°
50	20	30	diop	1178	plag in 1128°; ness in 1028°; B.M. 965 ± 5°
20	20	60	tridy	1363	diop in 1263°
75	15	10	ness	1233	diop in 1147°; B.M. 1035 ± 5°
65	15	20	ness + diop + plag	1118	B.M. 1020 ± 5°
55	15	30	diop	1148	plag in 1133°; ness in 1048°; B.M. 965 ± 5°
40	15	45	diop	1193	plag in 1112°; cristo* in 1035°; B.M. 1000 ± 10°
75	10	15	ness	1233	diop in 1111°; B.M. 1028 ± 5°
70	10	20	ness	1180	diop in 1098°; plag in 1067°; B.M. 1003 ± 5°
65	10	25	plag	1113	ness in 1103°; diop in 1083°; B.M. 1005 ± 5°
60	10	30	plag	1125	diop in 1093°; ness in 1058°; B.M. 965 ± 5°
55	10	35	plag	1143	diop in 1118°; ness in 1043°; B.M. 965 ± 5°
35	10	55	diop	1183	plag in 1073°; cristo* in 1063°; B.M. 1028 ± 5°
65	5	30	plag	1113	ness in 1083°; diop in 1052°; B.M. 985 ± 5°
60	5	35	plag	1133	ness in 1048°; diop in 1048°; B.M. 985 ± 5°
45	5	50	plag	1123	diop in 1098°; cristo* in 1035°; B.M. 1000 ± 10°
35	5	60	diop	1123	plag in 1073°; cristo* in 1073°; B.M. 1025 ± 5°
29	4	67	cristo*	1223	diop in 1128°
35	2	63	cristo*	1088	plag in 1070°

* Metastable phase at this temperature; tridymite is the stable phase.

Symbols: diop, diopside; ness, nepheline; fo, forsterite; tridy, tridymite; cristo, cristobalite; plag, plagioclase; B.M., beginning of melting.

(b) Mixtures on the line diopside-albite

Composition		Results of microscopic and/or x-ray examination		
Diopside	Albite	Primary phase	Liquidus temp.,°C	Other data
90	10	diop	1368	
80	20	diop	1347	
70	30	diop	1323	
55	45	diop	1293	plag in 1122°
40	60	diop	1263	plag in 1122°

25	75	diop	1226	plag in 1122°
12	88	diop	1158	plag in 1132°; B.M. 1043 ± 5°
10	90	diop	1140	plag in 1133°; B.M. 1043 ± 5°
8	92	plag	1138	diop in 1123°; B.M. 1045 ± 5°
5	95	plag	1148	diop in 1096°; B.M. 1028 ± 5°
2	98	plag	1143	diop in 1068°

(c) Mixtures on the line diopside-jadeite

Composition		Results of microscopic and/or x-ray examination		
Diopside	Jadeite	Primary phase	Liquidus temp., °C	Other data
50	50	diop	1253	
40	60	diop	1238	
30	70	diop	1203	B.M. 1005 ± 5°
25	75	diop	1180	B.M. 1005 ± 5°
20	80	diop	1155	plag in 1108°; B.M. 1000 ±5°
15	85	diop	1125	plag in 1113°; B.M. 985 ± 5°
10	90	plag	1111	diop in 1095°
5	95	ness	1113	plag in 1103°; diop in 1056°

compositions are in weight percent. Boundary curves are indicated by heavy lines, or by heavy dashed lines where their position is not accurately determined. The joins diopside-albite and diopside-jadeite are shown as light lines.

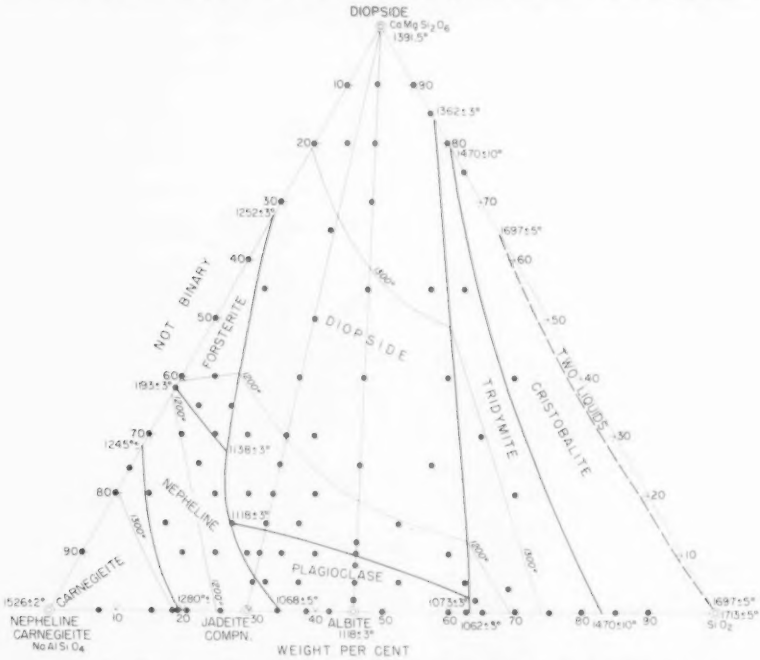

Fig. 2. Phase-equilibrium diagram of the system nepheline-diopside-silica.

Black dots represent the compositions studied and open circles the composition of compounds. Isotherms are given as light lines with the temperature indicated on each.

Data for the join albite-diopside are plotted in figure 3. Attention is called to liquidus temperatures for compositions rich in albite. From 1118°C, the melting point of pure albite, the liquidus rises to a maximum of 1148° and then falls to 1133° at the composition albite 91 diopside 9 where soda-rich

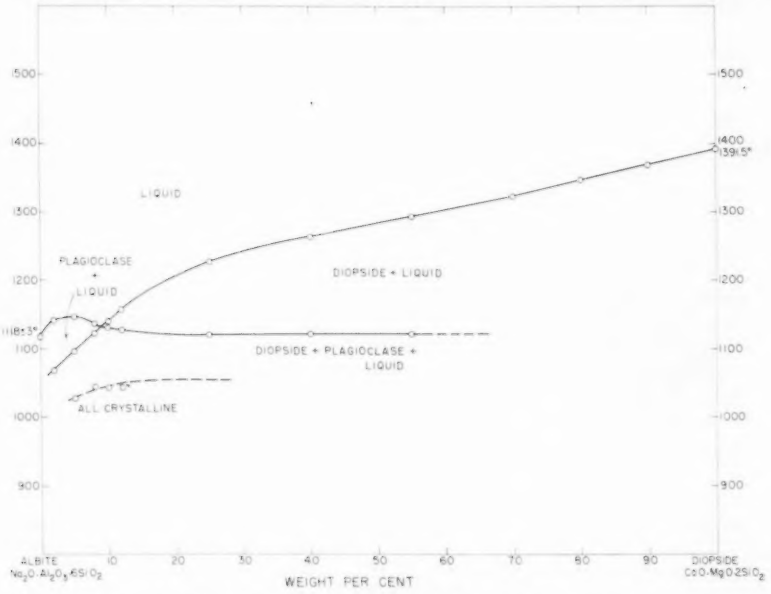

Fig. 3. Phase-equilibrium diagram of the join albite-diopside.

plagioclase and diopside coexist as primary phases on the liquidus surface. Unfortunately it is not possible to determine the compositions of the soda-rich plagioclases in the tiny crystals present in our melts by either optical or x-ray powder methods. If the diopside crystals were pure $CaMgSi_2O_6$ all compositions which have a diopside liquidus should be joined by the soda-rich plagioclase at the same temperature, 1133°C. An examination of the data in table 1 and figure 3 shows that as the diopside content increases diopside crystals are joined by soda-rich plagioclase at progressively lower temperatures. It is concluded from this evidence that there must be a small amount of solid solution in the diopside crystals. It is probably a slightly aluminous diopside. The x-ray powder patterns of the crystals show no significant difference from crystals of pure diopside.

In those mixtures in albite-diopside that were examined, only x-ray lines of sodic plagioclase and diopside were observed in completely crystalline mix-

tures. No evidence of any additional solid phase was observed. On the other hand, compositions in nepheline-diopside-silica more siliceous than those in albite-diopside showed the presence of silica (metastable cristobalite or stable tridymite) in the wholly crystalline mixtures, while those less siliceous than those in albite-diopside showed the presence of nepheline. This join albite-diopside is the thermal barrier or dividing line between compositions that yield siliceous residues on crystallization and those that produce nepheline-bearing residues.

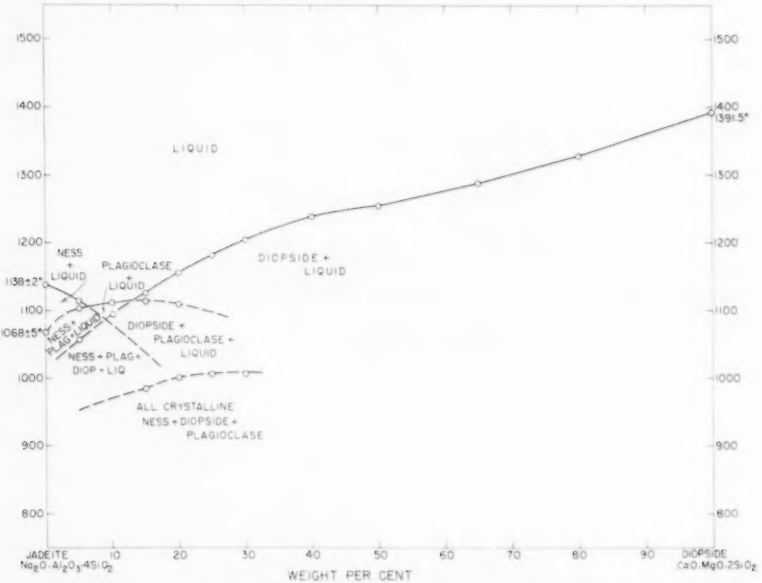

Fig. 4. Phase-equilibrium diagram of the join jadeite-diopside.

Data for the join jadeite-diopside are plotted in figure 4. An examination of this figure and the data presented in table 1 shows that there is nothing unique about this join at a pressure of 1 atmosphere. It cuts the liquidus surfaces of diopside (probably slightly aluminous diopside), soda-rich plagioclase and nepheline. X-ray powder patterns show that the nepheline is not pure nepheline ($NaAlSiO_4$) but a solid solution. Many of the x-ray lines of pure nepheline are displaced in the solid solutions. Compositions in the join jadeite-diopside, just like all compositions in nepheline-diopside-silica which are less siliceous than the join albite-diopside, consist of the three solid phases diopside, plagioclase and nepheline when they are completely crystallized.

Some compositions in nepheline-diopside-silica have forsterite as the primary phase. At lower temperatures forsterite disappears by reaction with liquid. No melilite was encountered at any temperature in mixtures in the system nepheline-diopside-silica, although it appears as one of the solid phases

at subliquidus temperatures in the system nepheline-diopside. In compositions in nepheline-diopside-silica near the side line nepheline-diopside where melilite and/or plagioclase might be expected, only nepheline and diopside were detected in the x-ray powder patterns of the completely crystalline mixtures. The lack of plagioclase or melilite may be due to inability of x-ray methods to detect small amounts of these phases. Where the mixtures are sufficiently away from this side line the x-ray patterns show plagioclase in addition to nepheline and diopside.

GEOLOGICAL SIGNIFICANCE OF THE SYSTEM

The results obtained support Bowen's thesis that through fractionation the magma is impoverished in the early-crystallizing minerals, in this case diopside, and enriched in other constituents, the ultimate goal being a liquid very rich in the alkali aluminosilicates. The exact nature of the residual liquids in this system is not known because the compositions of the liquids do not lie in the plane investigated. However, one very significant observation can be made with regard to the trend of the liquids. It was shown that for compositions in the triangle nepheline-diopside-albite the liquid became enriched in the nepheline component as opposed to a liquid arising from compositions in the triangle diopside-albite-silica which became enriched in the silica component. The join diopside-albite thereby becomes an "equilibrium thermal divide." It is clear that no single anhydrous composition can yield a liquid which through fractionation produces both a silica-rich residue and a nepheline-rich residue. This conclusion has important bearing on the nature of the parental basalt or basalts which produce a tholeiitic trend on the one hand and an alkali basalt trend on the other.

Because of the great importance of the diopside-albite join, it is now necessary to investigate further the nature of the residual liquids of the system albite-anorthite-diopside studied many years ago by Bowen (1915). At that time accurate temperature regulators were not available for long runs, and equilibrium could not be achieved in short runs because of the viscous behavior of the soda-rich liquids. New techniques may now be applied for which some degree of success can be anticipated in determining the course of crystallization of the residual liquids in that system. The residual liquids of the system resulting from the "plagioclase effect" (Bowen, 1945, p. 88) are of particular interest because of their possible bearing on the pantellerites and comendites, rocks related to petrogeny's "residua" system.

Another significant observation which may be drawn from the results of the nepheline-diopside-silica system is the reaction relation involving forsterite. It is not uncommon for some types of basalts to have modal olivine in excess of normative olivine. This is usually accounted for in the olivine tholeiites and tholeiites by failure of the olivine-hypersthene reaction. Similar reaction relations were thought to be absent in the alkali basalts (Wilkinson, 1956). It appears likely from the present work that at least one type of olivine reaction relation may also be found in the tephrites and related rocks such as basanites, melilite basalt, and possibly olivine nephelinites. The relation involves the reaction of forsterite with liquid to produce diopside. Other olivine reactions

involving melilite are known from the work on the nepheline-diopside join mentioned above.

The nature of the solid solutions in the investigated system has not been determined because an insufficient number of parameters are known to fix their complex compositions. For example, the proposed substitution of alumina into diopside may take place as in jadeite, $NaAlSi_2O_6$ or as in the alleged Tschermak molecules, $MgAl(AlSi)O_6$ or $CaAl(AlSi)O_6$. Jadeite was not observed by Greig and Barth (1938) in the nepheline-albite system nor by the writers in the range investigated for the present system. Jadeite is now believed to form only at very low temperatures when the pressure is low (Robertson, Birch, and Macdonald, 1957). It is of interest that the aluminous diopside fassaite, believed to contain a large amount of the Ca-Tschermak molecule, has not been found in quartz-bearing rocks, whereas jadeite is known to occur with quartz. The nature of the alumina in the diopside therefore may influence the course of crystallization, producing either a silica-rich residue or a nepheline-rich residue. The variability of composition of pyroxene may be able to provide the principal degrees of freedom required of a parental basalt.

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HOW OLD ARE THE ANGIOSPERMS?*

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ABSTRACT. The currently known paleontological evidence bearing on the time and appearance of the angiosperms in the fossil record is examined. Special attention is directed to the very considerable body of data derived from palynological studies made during the past decade in various parts of the world. It is concluded that no *bona fide* angiosperm remains, either megafossil or microfossil, have yet been described from rocks older than Early Cretaceous sediments. Their record, however, is extraordinarily and increasingly well-documented in post-Early Cretaceous sediments from the major geographic areas of the earth. Considering the efficiency of the atmospheric and fluvial media of transport of plant microfossils, and the well known instances of long distance transport of microfossils, it seems highly implausible that a major, evolving unit of the earth's terrestrial flora could have greatly antedated the Cretaceous without detection. The morphological problems in identification of "early" angiosperms and difficulties in determining their first appearance are briefly considered.

INTRODUCTION

The ancestry of the flowering plants and the place and time of their origin constitute a most conspicuous and also much discussed gap in paleontological knowledge. There is increasing agreement, in which we concur, that the original home of the angiosperms was in tropical rather than temperate extratropical regions, but the time of their differentiation is yet curiously uncertain. The rise and ultimate dominance of the angiosperms during the Cretaceous period is well documented in the geologic record despite the lack of fossil evidence regarding ancestral forms. The evolutionary advancement and diversity commonly attributed to Early Cretaceous representatives of the angiosperms has been interpreted to imply either (1) their relatively rapid evolution in middle Mesozoic time, or (2) their extended pre-Cretaceous existence.

Much of the recent speculation on the origin of the angiosperms has favored the second of these possibilities, postulating their existence as long ago as the Paleozoic (Axelrod, 1952, 1959; Camp, 1947; Eames, 1959; Lawrence, 1951; Thomas, 1957, 1958; Thompson, 1952). For example, Axelrod (1952, p. 57) summarizes his views on the early history of the angiosperms as follows:

> Geological and botanical data are adduced which support strongly the inferences that (1) angiosperms probably were in existence by Permo-Triassic time, that (2) they probably occupied upland regions remote from lowland sites of deposition, that (3) they were distributed in the diverse environments of the tropical zone, that (4) at least some of them were ancestral to modern primitive types of monocotyledons and dicotyledons which are now represented in tropical regions, and that (5) they were then undergoing quantum evolution from proangiosperms which occupied upland regions.

The present writers are not convinced that the available data indicate or even strongly suggest that the angiosperms originated as long ago as the Paleozoic. Therefore they have examined and here summarize some of the tangible evidence and inferences bearing on this problem.

PRE-CRETACEOUS RECORD OF THE ANGIOSPERMS

The only conclusive argument for the existence of pre-Cretaceous flowering plants would be concrete evidence in the form of fossil remains. A number

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of putatively angiospermous plant fossils have been described from Jurassic and older rocks, but the angiospermous affinities of most of these remains are, as Edwards (1955, p. 170) puts it, "vague and dubious." In some examples, the true relationships of the fossil material are not determinable owing to poor preservation or to the often unrecognized morphological parallelisms between flowering and non-flowering plants. In most cases, reinvestigation has shown the original assignments to have been in error, or revisions in stratigraphy have abated the situation. It is probably no overstatement to say that there has not yet been described a pre-Cretaceous plant whose angiospermous affinities are unequivocal.

Some of the most commonly cited examples of early angiosperms are discussed below:

1. *Homoxylon* Sahni

This genus, since renamed *Sahnioxylon* (Bose and Sah, 1954), was described in 1932 for Jurassic wood from India thought to represent a vesselless dicotyledon. Hsü and Bose (1952) later showed on the basis of gross stem structure that the material represents the secondary wood of a bennettitalean plant, probably of the *Williamsonia* complex.

2. *Suevioxylon* Kräusel

This poorly preserved specimen of fossil wood containing vessels having a ring porous arrangement, from the Brown Jura of Germany (Middle Jurassic), was described originally as angiospermous (Kräusel, 1924). Later, however, Kräusel (1950, p. 95) expressed doubt, noting that the determinable characteristics of the wood are also present in gnetalean plants.

3. *Eucommiidites* (Erdtman) Couper

Pollen thought to resemble that of the monotypic dicotyledonous family Eucommiaceae was described by Erdtman (1948) from shale of Early Jurassic age.

Couper (1956, 1958) has restudied this and other material of the genus. He concluded that the fossil pollen has gymnospermous affinities. Single specimens of pollen from Jurassic strata of the Caucasus, thought to resemble grains of the Angiospermae, have been recorded by Naumova (1939) under the form generic names *Tryptycha* "with three folds", and *Monoptycha* "with one fold." The three-folded form appears comparable to *Eucommiidites; Monoptycha* is based on a monocolpate grain which could be gymnospermous.

4. Pollen resembling *Castalia*, *Nelumbium*, and *Magnolia* (*Simpson*, 1937).

Simpson's material often accepted as authentically recording Jurassic angiosperms, has been re-examined by Hughes and Couper (1958). They show that all forms from the Scottish Middle Jurassic Brora Coal belong to gymnospermous taxa named since Simpson published his study. Hughes and Couper conclude that this evidence of pre-Cretaceous angiosperms must be discarded.

5. *Sporojuglandoidites* Vishnu-Mittre

This genus is based upon a single specimen preserved in chert from the Upper Jurassic of India. Vishnu-Mittre (1955) suggests that the grain, which

he regards as polyporate, is angiospermous and resembles pollen of the Juglandaceae. Preservation as shown by the figures is poor, and the pores are simple, unlike those of the Juglandaceae. Comparable openings, possibly due to degradative processes, crystal growth during diagenesis, or to fungal attack have been observed by us in undoubted spores of similar age.

6. *Poroplanites, Classopollis, Circumpollis, Canalopollis* Pflug

Pflug (1953) based these genera on Jurassic spores and pollen grains with supposed angiosperm-like germinal apertures. Krutzsch (1955) identified similar material from Liassic strata. Potonié (1954) and Couper (1955) have offered evidence that *Poroplanites* is a folded fern spore, and Couper believes that the other three genera have gymnospermous relationships.

7. *Furcula* Harris

This genus, of which Harris (1932) records some 50 specimens grouped in one species, is a curious leaf from Rhaetic deposits of east Greenland. Cuticular structure is known and is basically dicotyledonous. Stomata are of the syndetocheilic type, though this does not, as Harris implies, signify angiospermic affinity since the syndetocheilic condition prevails among the Cycadeoidales of the Mesozoic. The finer reticulate venation is strikingly dicotyledonous, though the coarse venation and over-all shape of the leaf are not. There seems no question that *Furcula* is of Rhaetic age since, according to Harris, it occurs in the same matrix with such typically Rhaetic species as *Ptilozamites nilssoni, Anomozamites minor*, etc. In the absence of any conflicting evidence, we suggest that *Furcula* be regarded as a member of the seed fern complex. The persistence of seed ferns well into the Mesozoic has been clearly established in the past few decades.

8. *Fraxinopsis* Wieland

These structures from the Rhaetic of Argentina originally were compared with winged fruits of *Fraxinus*. Kräusel (1950, p. 95) suggests that they are cone scales similar to *Cycadocarpidium* Nathorst.

9. *Propalmophyllum* Lignier

Palm-like leaves, or more restrictively, the terminal portions of the rachis of fan-shaped ridged leaves have been described as palms by Lignier. The specimens are of Liassic age and have been regarded by some authors (Seward, 1931) as evidence of the Jurassic antiquity of the Palmae. Lignier's studies were published in a somewhat obscure journal and have not been widely cited. Insofar as we know, Lignier's original specimens have not been reexamined. There is no doubt that his figured specimens appear palm-like, although their preservation, as indicated in the published figures, leaves much to be desired in making critical botanical asignment. As well as to the palms, they might also be compared to the juvenile leaves of some cycads, especially certain species of the genus *Cycas*. In Lignier's papers, his figured specimens of *bona fide* palms do not greatly strengthen his inferences regarding the affinity of *Propalmophyllum* with the palms. Until there is morphological evidence to the contrary, we feel that Lignier's material is most logically assignable to the Cycadaceae-Cycadeoidaceae complex of Mesozoic gymnosperms.

10. *Sanmiguelia* Brown

Leaves of a distinctly palm-like gross structure, preserved as impresssions on relatively coarse sandstones of Triassic age have been described as possible precursors of the palms (Brown, 1956). There is no doubt that Brown's material is the most arresting evidence of the megascopic remains of putative pre-Cretaceous angiosperms which has yet been described. The plaited, multi-ribbed leaves, in some cases attached to a very truncated rachis, are reconstructed as attached to a conical apical meristem. The restoration figured by Brown does not add significantly to the inferred relationship with modern palms. On the other hand, *Sanmiguelia* does not resemble any known early Mesozoic plant and its true affinity is certainly an open question. Unfortunately, no vestige of organic material is present in the matrix so that anatomical study is not possible. On the basis of the wide range of structure to be found within the cyadophyte line of vascular plants we suggest that the plant be regarded tentatively as a cycadeoid.

Regarding the fossil history of the palms, it should be noted that the megascopic remains of unquestionable palms are first found in the Upper Cretaceous. Extensive data exist, largely unpublished, which indicate that the earliest microfossil record of the palms is likewise Upper Cretaceous.

11. *Tetraporina* Naumova and *Triporina* Naumova

Under these generic names, Naumova (1950) described microfossils which she considers to be angiospermous from the Lower Carboniferous of the Moscow Basin. Similar forms were reported from the Lower and Middle Carboniferous of the western Don Basin by Teteryuk (1956), who notes that these forms range from Carboniferous through Cretaceous.

The Carboniferous alleged angiosperm pollen is difficult to evaluate because in most cases illustration is by drawing rather than by photomicrographs, and neither illustration nor text supplies the critical details of wall and pore morphology necessary to assess angiospermous relationships. The morphology and range of variation shown by described species of *Tetraporina* and *Triporina* bear a striking resemblance to the morphology of species of *Tetraëdron*, a unicellular green alga, in all important respects including shape, size, wall sculpture variation, and presence of thickenings and/or processes at the wall angles. Pores are absent in figures of many of the species assigned to *Tetraporina* and *Triporina*; where pores are shown, they could be analagous to the short, bifurcate processes present in some *Tetraëdron* species. *Tetraëdron* is represented in some fossil pollen assemblages (Florissant flora: Leopold, unpublished data; Stinking Water flora: Gray, 1956). Final assessment of the Russian Paleozoic forms must await an opportunity to examine the material.

The skepticism with which the listed pre-Cretaceous "angiosperms" are treated is not meant to imply that we reject the possibility that the ancestral angiosperms may have existed in Jurassic time, or indeed, earlier. We suggest only that since the fossil record is basic to paleontologic interpretation, extreme care should be taken in evaluating and accepting the actual evidence regarding so crucial an issue as the early record of the angiosperms.

UPLAND ORIGIN OF THE ANGIOSPERMS

To explain the paucity of angiosperms in the pre-Cretaceous geologic record, advocates of a Paleozoic or early Mesozoic origin for the group usually suggest that they arose and developed in upland areas of tropical regions far distant from the usual sites of deposition, it being assumed that the usual sites of deposition are in the lowlands. This inference has been made by several authors (e. g. Arnold, 1947, p. 334; Berry, 1945, p. 89; Seward, 1931, p. 366; Thomson, 1952). Three lines of evidence for the assumption have been stated by Axelrod (1952), whose views may be summarized as follows:

1. The majority of plant fossils accumulate at or close to the site of deposition, typically in lowland lake or streamside regions of sediment accumulation. Since upland sites of deposition, when they occur, are not likely to persist for long periods of geologic time, the probability of finding fossil leaves representing older, upland angiosperms is slight or non-existent.

2. Cliseral relations demonstrated by successions of fossil floras show that higher and more specialized types were present in uplands of a region long before they entered the lowlands as replacements for the older flora.

3. Diversity of physical environment in upland areas would promote rapid evolution in the ancestral angiosperm types.

Discussion

Let us consider these three lines of evidence in more detail.

I. The conclusion that fossil plants represent chiefly vegetation that grew at or near the site of deposition is based primarily upon the occurrence of fossil leaves which, in the past, have been the most studied organ of extinct plants. Other plant organs of a less fragile nature, e. g. wood, certain fruits and seeds, and pollen may withstand transport for long distances; in fact, paleoecological interpretation of these remains must be made cautiously because of the likelihood that organs representing plants from diverse habitats have been assembled.

Examples of the transport of wood from higher environments to a lower environment of deposition include the principally Devonian genus *Callixylon*, considered by Axelrod (1952) to represent an upland gymnosperm, and also other associated woods in the New Albany shale (Hoskins and Cross 1951); *Araucarioxylon* assembled in logpiles in the Petrified Forest National Monument in Arizona and elsewhere on the Colorado Plateau (Daugherty, 1941); araucarian cones in great quantities in the petrified forest of Cerro Cuadrado, Argentina (Wieland, 1935); wood of *Ginkgo* and *Betula* associated, somewhat incongruously, with endocarpal remains of tropical lianas of the families Icacinadceae and Menispermaceae in the Clarno formation of Oregon (Scott, unpublished data); the Aptian dicotyledonous woods described by Stopes (1912); seeds and fruits of unquestionably upland habitat associated with remains of swamp plants in the rich flora of the Brandon lignite of Vermont (Barghoorn, 1950, and unpublished data).

However, in the record of upland vegetation preserved in sedimentary basins, it is microfossils rather than megafossils which are most likely to be encountered. Although the representation of wind-borne pollen in sediments is determined by the interaction of a complex of factors of the physical and biological environments (Leopold and Scott, 1958; Muller 1959), a pollen assemblage typically contains forms from diverse ecological situations. The large numbers of pollen grains produced, together with their suitability for preservation under a variety of geologic conditions make it statistically far more probable that pollen will be incorporated in sediments than leaves or other organs. The small size and buoyancy of most pollen make it especially suitable for distribution by either wind or moving water.

Examples of long distance transport of modern pollen by wind are common. Grayson (1960, oral communication) discovered *Quercus* and *Acer* pollen in several moss polsters from Labrador. He concludes that these occurrences represent wind transport from deciduous forests hundreds of miles to the south. The most extraordinary example is that of Hafsten (1951) who reports *Nothofagus* pollen in a modern peat profile from the islands of Tristan da Cuhna in the center of the south Atlantic basin. Geographic relationships and effective wind directions indicates that the *Nothofagus* grains were transported from southern South America, a distance of 4500 kilometers. Pollen of *Podocarpus* from upland habitats on the island of Trinidad has been transported by wind to marine sites as far as 75 miles from the island (Muller, 1959).

In fossil floras, Kosanke (1950) suggests that winged gymnospermous pollen of the genus *Illinites* in Pennsylvanian coal swamp deposits may have originated from plants growing in uplands. Study of the pollen from the Eocene Green River formation (Wodehouse, 1933) added several upland genera to the warm-temperature streamside assemblage known from leaves. Traverse (1955) identified pollen of such upland genera as *Pterocarya* and *Quercus* (in part) in the dominantly swamp assemblage of the Brandon lignite of Vermont. *Podocarpus*, now limited in Venezuela to mountainous sites above 2000 meters, occurs sporadically as pollen in sediments of Eocene age and younger, and increases markedly in sediments of Oligocene and Miocene age (Kuyl, Muller and Waterbolk, 1955) in Venezuela.

As noted above, water is likewise an effective medium for transport of pollen. Because of their small size, equivalent to silt particles, and extreme resistance to chemical and microbiological degradation, pollen grains may be carried for long distances as a part of the sediment load of streams before incorporation in lowland flood plains or even in marine deposits. Muller (1959) in his intensive study of pollen distribution in sediments of the Orinoco delta, shows by specific examples that, depending on the various factors, the pollen obtained in a single dredge sample may have been derived from cosatal vegetation bodering the estuaries or transported by air or water from inland montane environments. An example of long distance transport by water is the occurrence of *Alnus* pollen, which must have been carried more than 500 miles from mountain habitats by the Orinoco system.

Objection to the likelihood of the pollen of early upland angiosperms

entering the sedimentary record has been raised on the ground that they may have been pollinated by insects rather than wind. The geologic history of Coleoptera dates from the Permian, and the existence of beetle pollination in the woody ranalian complex has led some investigators to link their rise and to suggest co-existence (Grant, 1950; Eames, 1959). Modern members of the Magnoliales known to be pollinated by beetles, e. g. *Magnolia, Illicium, Calycanthus*, have pollen that is also well suited to transport by wind, being of small to medium size and produced in numbers. Ranalian floral characters, particularly the numerous stamens borne on an open flower and the medium to small sized pollen, which is within the range (17-58μ) of the majority of anemophilous plants (Dyakowska and Zurzycki, 1959), favor the representation of their pollen in the regional pollen rain regardless of the functional vector in pollination.

The absence or paucity of pollen of some ranalian genera in recent (i.e. post-glacial) sediments, including peat deposits, may be ascribed to lack of resistance to chemical and microbiological factors in the depositional environment. Thus pollen of the Lauraceae and many genera of the Magnoliaceae are quite ephemeral in modern sediments owing to feeble development of their exines and, more important, the chemical instability of the exine to oxidative and microbial attack. The pollen of many of the more primitive surviving ranalian forms, however, are featured by massive exines, and the grains with stand rigorous treatment with a variety of reagents as may be demonstrated in laboratory preparations.

That insect or bird pollinated plants may indeed have their pollen preserved in sediments is demonstrated by the fossil record of the Proteaceae in Australia, where Cookson (1950) has found a variety of grains assignable to this family in a number of Tertiary deposits.

It is unlikely that pollination in the primitive angiosperms was limited to insect vectors. Stebbins (1951) has pointed out that various means of pollination are utilized within most major taxa. He states that (p. 299): " . . pollination by flies, birds, wind and water is likewise scattered through the angiosperms with little relationship to the boundaries of the higher categories".

Thus for both the largest and the smallest of angiosperm organs, their woody trunks and their pollen, there exist prevalent, competent and demonstrated means for the representation of upland plants in lowland sediments. Emphasis on palynology in the last decade has resulted in the examination of thousands of samples of sediments of Mesozoic age from the major geographic regions of the earth including many from the tropics; but authentic pre-Cretaceous angiosperm pollen has not yet been reported. Although the leaves of upland dicotyledons cannot ordinarily nor logically be expected in lowland deposits, there is ample reason to give careful scrutiny as to why their pollen is consistently absent from pre-Cretaceous sediments if the group had actually been evolving in the uplands since the Paleozoic.

II. In cliseral relationships among floras, migration is by no means limited to the downslope path of movement, and the source and nature of the replacing flora is ecologically determined largely by the nature of the climatic

change. Cooling would cause upland, more temperate plants to move into lower latitude lowland regions; a widespread and consistent warming trend provides an opportunity for lowland plants to extend their ranges at the expense of the more temperate plants previously occupying the area. The first situation is illustrated by the southward retreat of the warm temperate and subtropical flora of late Eocene and early Oligocene time in temperate regions of Europe and North America; the converse is shown by the northward movement in the early Tertiary of elements of the Poltavian tropical or subtropical flora to the latitude of present-day southern England (Chandler, 1954; Edwards, 1955).

The fossil record provides exceptions as well as agreements to the thesis that higher, more specialized types of plants were present in the uplands of a region long before they entered the lowlands as replacements for the older flora. The first vascular plants, the Psilophytes, almost certainly arose from aquatic or semi-aquatic environments to dominate and replace the prostrate non-vascular land flora of the earlier Paleozoic lowland basins. There is no evidence nor implication in the fossil record that either the early or advanced arboreal members of the Lycopsida and Sphenopsida entered the lowland environment from the uplands to become a part of the dominant forests of the Carboniferous. In fact, their anatomical structure suggests the opposite. In the case of more highly specialized groups of plants of known Paleozoic origin, the Coniferales are of much interest in connection with the hypothesis that evolutionary specialization is fostered by high altitudes in tropical environments. The ancestors of the entire group of coniferophytes (including probably also the Ginkgoales), the Cordaitales, were among the most abundant representatives of the lowland Carboniferous swamp forests. It seems in accord with currently known facts of the fossil record that the major units or phylogenetic derivatives of the Cordaitales had diverged by the early Mesozoic; most probably the taxonomically recognizable families of the Coniferales were morphologically distinct by Jurassic time. It may well be argued that the Coniferales represent a major and, in the existing flora, a well-represented group of plants whose origin has been in lowland tropical regions and whose diversification has been in upland and cool temperate regions. In this connection it is of interest to note their present phytogeographic and evolutionary status. The conifers are a moribund group of plants in terms of the late Mezozoic and Cenozoic history of the earth's vegetation. Presently, their phyletic as well as geographic distribution is featured by a high degree of endemism, especially in higher altitudes of warm temperate and tropical regions. In fact, 15 genera of conifers are each represented in the surviving flora by only one species, and these species are chiefly limited to isolated upland localities in austral tropical areas. It would appear that the geographic origin and phylogenetic diversification of the Coniferales has been one of origin in lowland environments and progressive moribundity in upland environments, chiefly in tropical regions.

If the primitive angiosperms had become established in, and remained restricted to, uplands during all or a major portion of the time interval from the Permian to the Cretaceous, a widespread, major climatic deterioration presumably would have been required to initiate their invasion of the tropical

lowlands to displace the established gymnosperm flora on the extensive scale that occurred in the Early Cretaceous. Evidence from other paleoecologic considerations and groups of organisms (e. g. marine invertebrates: Gignoux, 1950, p. 392; and vertebrates: Colbert, 1953, p. 265) indicates that climatic conditions then, while not so uniform as in the Jurassic, were equable over large portions of the earth during Early Cretaceous time. If minor or localized ecologic changes were adequate to initiate the migrational capacities of the primitive angiosperms, it would be remarkable that this group of plants should be restricted geographically for such an extended period of geologic time as elapsed between the late Paleozoic and the Early Cretaceous.

III. There is little question that, in general, the variety of physical environments existing in uplands provides a favorable setting for rapid evolution and diversification (Simpson, 1944; Stebbins, 1950). The flowering plants also have found the tropical lowlands favorable for diversification; floras of the tropical rain forests are probably the richest plant assemblages known (Good, 1947, p. 192). The great variety of plants and their vast speciation in the tropics may be related to the fact that the majority of arboreal dicotyledons are basically similar vegetatively; important differences among them are more likely to involve aspects of reproductive structures and adaptations for seed dispersal. The presence of rich, rapidly evolving faunas of insects or other animals might well have been more fundamental to differentiation of taxa than stresses of the physical environment in the early stages of angiosperm evolution (Stebbins, 1950, p. 502). For example, some members of the Hymenoptera and Diptera, insect orders containing many forms now effective in flower pollination, existed in Jurassic time (Carpenter, 1955). The diversification of the angiosperms may be linked, at least in part, to the rise of these groups.

Little of the known background of the angiosperms relates them to temperate upland situations. On the contrary, there is much evidence linking the group to lowland, tropical conditions (Bews, 1927). Axelrod (1952), in summarizing Bews' work, comments that: "The evidence reviewed by Bews makes it seem reasonably certain that the phylum was basically adapted to moist tropical forests, and that it has become adapted to the more specialized environments of the tropical and extratropical regions".

This evidence, in extremely skeletonized form, includes in part the following: members of the ranalian complex, including the most primitive extant angiosperms, are primarily adapted to the environment of the moist tropical lowlands. Within the major taxa of flowering plants, the more primitive floral types are most commonly present in tropical environments, the derived types are associated with more specialized tropical and extra-tropical environments. Based on modern distribution patterns, three-fourths of all angiosperm families apparently are primarily adapted to tropical regions. It may be noted in this connection that the pollination mechanism of incompletely closed carpels, which yet exists among the most primitive surviving angiosperms (Bailey and Nast, 1943; Bailey and Swamy, 1951) would seemingly require the atmospheric environment of the humid tropics for effective operation. The persistence in the humid tropics of the primitive vesselless dicotyledons of the ran-

alian families Winteraceae, Amborellaceae and Sarcandraceae is therefore quite in accord with their presumed morphological dependence on a humid tropical environment.

Studies of changes in modern tropical vegetation with increasing altitude have been summarized by Richards (1957, p. 346) who states that: "The tall luxuriant Tropical Rain forest of the lowlands gives place to other communities also evergreen, but lower in stature, simpler in structure, and floristically poorer. The purely tropical flora is left behind and is replaced by a montane flora in which many of the genera, or even the species, are temperate." Five major zones of vegetation, often sharply delimited, can be distinguished. The general features of this altitudinal zonation appear similar in both Old and New World tropics, and must have existed in the past.

The existence of this zonation suggests that, had the early angiosperms been restricted from the Permian to the Cretaceous to upland habitats of sufficient altitude to furnish environmental diversity of effective contrast to that of the tropical lowlands, the resultant flora would have developed a more temperate aspect rather than the lowland character to which the available evidence points.

SOME ASPECTS OF EARLY ANGIOSPERM FLORAS

In marked contrast to the uncertainty surrounding their existence in pre-Cretaceous rocks, there is positive and widespread evidence from Cretaceous sediments of the entry and increase of angiosperms in the fossil record. Leaf floras documenting the transition from dominantly gymnospermous to dominately angiospermous plants have been compiled by Axelrod (1959, p. 203), who interprets them to show that "flowering plants first entered the lowland record within the tropical belt at lower middle latitudes at the beginning of the Early Cretaceous; that they reached high latitudes late in the epoch, and that, with few exceptions, they became dominant there only in Late Cretaceous time."

There is further evidence based on palynological studies of Mesozoic sediments which bears out the genuineness of the Cretaceous entry of the flowering plants. As pointed out earlier, fossil pollen is especially pertinent evidence as to the time of the first appearance of the angiosperms because of the likelihood that it will record upland as well as lowland vegetation.

Angiosperm pollen has been found to be lacking in Late Jurassic and earliest Cretaceous sediments in both eastern and western hemispheres. Quantitative studies of pollen currently known from Cretaceous sedimentary sequences, however, consistently record the appearance and increase of the angiosperm population and their ultimate dominance.

One study is that of Couper (1958), who examined 100 samples from Jurassic and Lower Cretaceous Wealden and Aptian beds in Great Britain and Scotland. He found the Jurassic and Cretaceous floras to be distinct, but noted no definite angiospermous pollen grains in either. Delcourt and Sprumont (1955) found no angiosperm pollen among plant microfossils from Wealden beds in Belgium. Angiosperm pollen is absent from a rich assemblage in the Morrison formation, of Upper Jurassic age in Utah (Scott, unpublished

data). A survey of some 20 pollen-bearing samples of Triassic and Jurassic age from various localities in Europe, North America, and South America has failed to turn up any angiosperm pollen (Scott, unpublished data). Pollen and spore assemblages from the Upper Triassic Keuper of Germany (Leschik, in Kräusel and Leschik, 1956) and in the Upper Triassic Chinle formation of Utah and Arizona (Scott, unpublished data) lack angiospermous pollen as does also a Jurassic pollen flora from India (Vishnu-Mittre, 1954). Examination of microfossil floras of Jurassic and Upper Triassic age from such diverse areas as Greenland, equatorial Africa, Sweden, Great Britain, Central Europe and eastern North America have likewise failed to reveal any evidence of the existence of angiosperms (Barghoorn, unpublished data).

Palynological examination of sequences of Cretaceous sediments ranging from the lower Trinity (Aptian) through the Navarro (Maastrichtian) and into the Midway (Paleocene) in central Texas has been made by Grayson and Pierce (1959, personal communication, and also communicated to the 9th International Botanical Congress, Montreal, 1959). Their work shows a progressive increase in both the absolute and relative number of angiosperms represented, from their absence in the lowest beds to a predominance (75% of the total pollen and spores present) by the end of Cretaceous time. Krutzsch (1957) records that typical angiosperm pollen is not present in Middle European sediments until Albian and Cenomian time. Angiosperm pollen first apepars in the fossil record of New Zealand in Cretaceous time (Couper, 1953). In a detailed palynological study Steeves (1959) has found that in the Cretaceous sediments represented on Long Island, New York, there is a progressive increase in the representation of angiosperm pollen from the lowermost (Raritan age) beds to the later (Magothy age) sediments. Concomitant with the increase in angiosperms there is a progressive decline, both in absolute and relative proportions, of gymnosperms and ferns. The data presented by Steeves (1959) are based on samples secured from continuous subsurface cores extending from below the drift to the bedrock (saprolite, of unknown age) on eastern Long Island. The borings penetrated approximately 1400 feet of non-marine Cretaceous sediments and hence constitute a representative sample of the known middle to Upper Cretaceous section of the coastal plain of eastern North America.

A factor in the extrapolation of the angiosperms far into the geologic past is the belief that the forms found in Early Cretaceous rocks were highly evolved representatives of modern families or even genera; hence, they must necessarily have had an extended past history. For the most part this conclusion is based upon the identification of fossil leaves. The problems involved in the identification of detached leaves have been discussed by Bews (1927 p. 4-5), Good (1947, p. 264-265) and Stebbins (1950, p. 515-516). The latter author concludes that "forms similar to modern species and genera are recognized with relative ease, while radically different types, including those which might have formed connecting links between existing genera or families, could not be recognized or assigned to their correct phylogenetic position even if they were found." Stebbins (1951, p. 301) further points out with reference to floral structures the likelihood " - - that in the remote period of geologic

time, probably the Cretaceous period, when the ancestors of the modern families became differentiated from each other, their interrelationships were much like those between modern species and genera, and were governed by similar selective factors". This conclusion would seem to be applicable to leaf as well as floral morphology, and points up the possibly misleading nature of many early dicotyledonous leaves. If there are criteria by which fossil leaves from the Early Cretaceous can be assigned reliably to modern families, they have not been delineated in the literature despite the many definite assignments that have been made.

The aura of modernity surrounding the Early Cretaceous dicotyledons is created in no small part by the practice of erecting names for form genera based upon supposed resemblances to modern genera, *e.g. Quercophyllum*, and the use of modern generic names for leaves which would better be placed in form or organ genera. An example of the effect of this practice is the common, uncritical acceptance of the increasing percentage of angiosperm species in successively younger beds of the Lower and Upper Cretaceous Potomac formation of Maryland as evidence for the rapid rise of the angiosperms at this time. Although forms in the lower beds of the Patuxent and Arundel have been given such names as *Juglandiphyllum* and *Ficophyllum*, there is reason to doubt whether most of the leaves are even angiospermous (Berry, 1911).

The dicotyledonous woods from the Aptian beds of England are another example pointing to the supposed modernity or high level of specialization of Early Cretaceous angiosperms. The Aptian woods were described before many of the criteria for phylogenetic advancement of the secondary xylem of woody dicotyledons were formulated. Bailey (1949) has observed that the full range of morphological expression from primitive to advanced can be found among the extant dicotyledons. The Aptian woods exhibit much of this range; it is not surprising therefore that the Aptian woods have a "modern" aspect. The possibility exists, unfortunately, that the specimens described by Stopes were from beds other than Aptian since they were selected from museum specimens which may or may not have been authentically placed stratigraphically (Couper 1958). There is need for further study of Aptian age beds in England before the significance of the Aptian woods can be established firmly. As previously noted, angiosperm pollen is absent from the Aptian beds of Great Britain (Couper, 1958).

There is morphological evidence from reproductive structures which indicates that Cretaceous angiosperm floras may have had a much less modern aspect than is commonly attributed to them. For example, Cretaceous pollen assemblages, particularly those from the early Late Cretaceous, contain a wealth of angiosperm forms whose detailed pollen morphology is unlike that of known living plants (Steeves, 1959). In many cases pollen cannot be assigned to families; where families can be recognized, modern generic assignments may be difficult or impossible. Although there is great variety in wall and aperture morphology in Cretaceous angiosperm pollen, grains representing certain major morphological classes do not appear until the Tertiary (Kuyl, Muller and Waterbolk, 1955).

Angiosperm fruits provide highly reliable guides to systematic affinities.

An angiosperm fruit from the Lower Cretaceous of France could not be related by Chandler (1958) to any modern genus or family. Chandler (1954) notes that Upper Cretaceous fruits from Egypt are difficult to relate to modern genera and families. Even as late as the early Tertiary, fossil reproductive structures, where known, show that the modernity of the floras was diluted by a significant admixture of extinct forms (genera). In the lower Eocene London Clay flora, Reid and Chandler (1933) found 75% of the genera to be extinct. Extinct genera represented by fruits and seeds have been recognized in the Tertiary brown coals of Germany (Kirchheimer, 1957), in Clarno flora (Eocene) of Oregon (Scott, 1954, 1956) and in the Brandon lignite (Oligocene?) of Vermont (Barghoorn, 1950; Barghoorn and Spackman, 1950).

Although the angiosperms constitute the most highly evolved group of vascular plants, many of the morphological features which characterize them also developed in other groups. Thus vesselless wood indistinguishable from that attributed to primitive angiosperms existed in the extinct cycadophytes (Hsü and Bose, 1952); even vessels, which characterize most modern angiospermous woods, have evolved independently in six groups of vascular plants (Bailey, 1949; Bierhorst, 1958). Leaves having reticulate venation are known in diverse groups, e.g. Gnetales, ferns, pteridosperms. The flower is difficult to define in a manner that excludes the highly organized inflorescenses of the cycads, and angiospermy was approached in the Caytoniales and in early Mesozoic seed ferns. Only double fertilization and endosperm development seem to be truly unique angiospermous achievements, and for the origin and significance of this the fossil record can provide no clue.

Until the angiosperm ancestors are identified, there remains uncertainty as to the extent of the morphological gaps between the flowering plants and their predecessors. The occurrence, due to parallel development, of many of the advanced features of the angiosperms in separate taxa of vascular plants raises the possibility that, whatever the relationships of these elusive ancestors, they had already evolved in their structural organization features now characteristic of the angiosperms. If several of these features, e.g. vessels, reticulate leaf venation, and bisexual reproductive organs, had already evolved in precursors, the evolutionary steps that differentiated the flowering plants may not have been great, and the geologic time required for their diversification not extreme. In this connection it is of interest to note the extraordinary diversification and ubiquitous distribution of herbaceous dicotyledons whose evolution has apparently been a phenomenon of post early Miocene time, as deduced from abundant palynological evidence.

SUMMARY

A number of lines of evidence suggest that there may be a shorter interval of geologic time between the origin of the angiosperms and their first appearance as fossils in the Cretaceous than is allowed by any theory postulating their existence in tropical uplands since the late Paleozoic. This evidence includes the lack of unequivocal angiospermous plant fossils in rocks of pre-Cretaceous age and a consistent record, based upon pollen as well as leaves and other organs, of their entry into and rapid numerical increase in Lower Cre-

taceous sediments. The common intermingling in sediments of pollen derived from plants growing in diverse habitats, including uplands, makes it quite unlikely that angiosperms could have persisted in upland regions from the Paleozoic to the Cretaceous without having contributed microfossils to sedimentary basins. Additionally, known primitive angiosperms as well as a majority of all angiosperm families basically are adapted to equable tropical conditions rather than to more extreme, upland environments.

The time of the origin of the angiosperms, whether Paleozoic, Cretaceous, or, as is more probable, intermediate between them, is not known. The fascinating potentialities of the paleontological record make it possible and even likely that tomorrow's discoveries will outmode today's speculation. Despite its vagaries and imperfections, the fossil record remains our best index to relationships involving geologic time; and it does not bear out speculations on the origin of the angiosperms in the Paleozoic era.

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NOTES ON THE MEASUREMENT OF FAUNAL RESEMBLANCE

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ABSTRACT. Of various measures of taxonomic resemblance, the percentage in the smaller of two faunal samples of the number of taxa common to both is most useful. It tends to eliminate the effects of discrepancy in size between the two faunas or samples, and when that discrepancy is a factor in the problem being studied the percentage of common taxa may be used. When faunas closely similar taxonomically are compared, it may be desirable to take into account differences in the relative abundances of taxa in common. For that purpose, measures based on rank correlation are suggested, but no such measure seems full satisfactory. Several other measures used in current literature are discussed.

MEASURES BASED ON NUMBERS OF TAXA

Most studies of faunal resemblances are strictly zoogeographical or correlational (in a stratigraphic sense), although other and especially ecological factors cannot be wholly excluded and are sometimes of primary importance. The usual approach is concerned with the presence or absence of taxa and with degrees of phylogenetic relationships among them. Degrees of phylogenetic affinity are practically, even though somewhat crudely, quantified by the levels of hierarchic classification. Comparisons at the specific level involve close affinity, at the generic level broader affinity, at the family level still broader, and so on. Presence and absence of taxa, for purposes of measurement of faunal resemblances, are adequately quantified by the following symbols and concepts:

E_1, Number of taxa (at a specified level) in the first, smaller (or equal) of the two faunas or samples compared, but absent in the second.

E_2, Number of taxa in the second, larger (or equal) fauna or sample but absent in the first.

C, Number of taxa common to both.

$N_1 = E_1 + C$, Total taxa in first.

$N_2 = E_2 + C$, Total taxa in second.

$N_t = E_1 + E_2 + C = N_1 + N_2 - C$, Total taxa in both.

The most obvious, and apparently the most intuitively acceptable measure of faunal resemblance, is simply the percentage of taxa in common among the total taxa of the two faunas or samples in question:

$$\frac{C}{N_t} \times 100 \qquad (1)$$

which for purposes of compilation and calculation is more conveniently represented as:

$$\frac{C}{N_1+N_2-C} \times 100 \qquad (1a)$$

If both faunas are almost or quite completely known, and if they are of at least approximately equal size, index (1) is usefully indicative. The maximal value, 100, shows complete indentity and the minimal, 0, complete difference at the chosen taxonomic level; and, in general, intermediate values

are clearly proportionate to degrees of resemblance. If, however, either fauna is incompletely known, or if the known numbers of taxa are distinctly unequal index (1) is difficult to interpret and may be quite misleading.

It is generally the case with fossil faunas and sometimes with recent faunas that the available samples, on which a measure must be based, are incomplete and do not include all the taxa of both faunas. If the two samples have the same degree of completeness relative to the populations[1] (e. g. each has half the taxa of the respective population) and neither is biased with respect to E/C, index (1) from the samples may give comparable estimates of the population values. The actual ratio of sample N to population N is, however, always unknown and can rarely be closely estimated. The best one can do is to assume, in the absence of indications to the contrary, that the two populations have about the same numbers of taxa. If that assumption is true, comparisons of samples of approximately equal size still give somewhat biased estimates of index (1), but estimates that are reasonably comparable and are adequate if the samples are fairly large.

Those conditions are, however, rarely met. Samples are frequently of quite unequal size. It is frequently probable that the smaller sample (smaller in terms of number of taxa included) also has a smaller proportion of the taxa originally present in the corresponding population. Then the value of index (1) will be especially strongly biased as an estimate of the population value, and the bias will be greater the greater the discrepancy in N_1 and N_2 in the samples. The mere fact that N_2 is larger makes index (1) smaller. A quick and intuitively clear way to check such bias, for this or any other index, is to consider samples drawn from indentically the same population. If bias is absent, they should of course tend to give the maximal value of the index. It will at once be seen that index (1) cannot have its maximal value, 100, if the samples are of unequal size and that it may give extremely low values for samples from the same population. Obviously, comparisons based on this index are usually unreliable for unequal faunas or samples.

The following index eliminates the worst disadvantages noted for (1):

$$\frac{C}{N_1} \times 100 \qquad (2)$$

As an estimate of a population index from samples, index (2) minimizes effects of differences in size between N_1 and N_2. When samples are small, both (1) and (2) of course have considerable sampling error; but (2) is also preferable in this respect, and indeed the larger the discrepancy between N_1 and N_2 (the better N_2 but not N_1 is sampled), the lower the sampling error without, as with (1), introduction of increasing bias. When the sample (and

[1] Here and elsewhere in this paper, the word *population* refers to the actual fauna under consideration as it exists in nature. It might, for instance, be all the native mammals now present in a defined area, or all the fossil pelecypods preserved (but not necessarily collected) in a given stratum. When taxonomic indices are being used, the pertinent population is all the *taxa* (at the level involved in the index) present in the defined fauna in nature. Sampling error is involved when any of those taxa are unknown. When indices take individual abundance into consideration (discussed later in this paper), the pertinent population is all the individuals in the defined fauna. Then study is virtually always on a sampling basis.

population) sizes are indeed equal, (2) is still at least as good an estimate as (1), so there is no reason to turn to (1) in this (unusual) situation.

Note that (1) and (2) both have scales from 0 to 100 but that they do not tend to be equal except when (1) is 100 or either one is 0. [(1) is not, however, necessarily 100 when (2) is 100.]Otherwise (2)is always somewhat larger than (1). These, like any two different indices, measure different things, and values are to be compared only when they represent the same index.

When both faunas are completely known taxonomically, the comparison is on the true population basis rather than being based on incomplete samples. This situation is common in comparing Recent faunas of higher vertebrates, at least, but rarely arises in paleontology. When the populations are completely known, the advantages of (2) may be less obvious but may still exist. If the numbers of taxa in the two populations are equal, both indices make consistent (not identical) comparisons and there is no particular reason to prefer one over the other. If difference in sizes is in itself a meaningful element in the desired comparison, then (1) is preferable. Other things being equal, index (1) but not (2) will be smaller the greater the discrepancy between N_1 and N_2, and this is desirable if the discrepancy is important for the particular problem being investigated.

That is not, however, always or even usually the case. If, for instance, comparison is of the fauna of a smaller with that of a larger area, the zoogeographic relationships are more clearly indicated if it is possible to eliminate the tendency of larger areas to have larger faunas simply because the areas *are* larger, and then index (2) is clearly more indicative. Index (2) is even more obviously advantageous when the smaller fauna has been derived from the larger, or when a local fauna is compared with a regional fauna. The situation is comparable in stratigraphic correlation when, as is so often done, a particular faunule is compared with the whole known fauna of a given age. In these or other comparisons, it also frequently is true that the larger fauna is more varied ecologically. Then use of index (2) tends to minimize the merely ecological difference, although obviously this cannot be wholly eliminated. In effect, index (2) tends to stress the most nearly similar parts of the two faunas, which is usually an advantage in zoogeographic and especially in stratigraphic investigations. Finally, since index (2) is generally less biased for paleontological samples, it is advantageous to use (2) consistently and thus to permit more nearly valid comparisons of both fossil and recent faunas. (For further discussion and for extensive exemplification of index (2) see, e.g. Simpson, 1947).

Burt (1958) suggests that what is here called index (2) "should be applied in both directions to give the true picture," in other words that one should give not only

$$\frac{C}{N_1} \times 100 \qquad (2)$$

but also

$$\frac{C}{N_2} \times 100 \qquad (3)$$

This is not, in fact, the application of the same measure in two directions, but the addition of another measure with different properties. Unless the samples are equal in numbers of taxa, when (2) and (3) become identical, and except at the extreme values 100 and 0, (3) is always smaller than (2) for the same comparisons. If one thinks of these indices as truly analogous or as comparable with each other, it may appear that fauna 2 resembles fauna 1 less than fauna 1 resembles fauna 2, which is confusing. In fact (3) is a distinctly different measure that suffers to exaggerated degree from the disadvantages of index (1), already discussed. If the discrepancy in sizes of samples is considered significant and is wanted to influence the index, then (1) is available and is more readily understood and more widely used. Otherwise (2) is better. There seems little reason to use index (3).

Still another index that is a sort of average (but not the arithmetic mean) between (2) and (3) has been advocated especially by Pirlot (1956; see also Burt, 1958, and Crusafont and Truyols, 1958):

$$\frac{2C}{N_1+N_2} \times 100 \qquad (4)$$

This is the percentage of common taxa not in the total for both samples as in (1), or in one sample or the other as in (2) and (3), but in the mean number of taxa for the two, as is more obvious when the formula is written:

$$\frac{C}{(N_1+N_2)/2} \times 100 \qquad (4a)$$

This is in effect a compromise between (1) and (2), for it reduces the effect of discrepancies in sample size, which is prominent in (1), but does not minimize that effect as does (2). It is hard to see a meaningful use for (4), because when discrepancy in size is pertinent to the problem (1) is better, and when it is not, (2) is better.

In contrasting (1) and (4), Burt (1958) notes that in (1) C appears in both numerator and denominator and in (4) only in the numerator. That is, however, misleading because C is really present in both numerator and denominator of all the indices (1)-(4), as is evident when they are rewritten as follows:

$$\frac{C}{E_1+E_2+C} \times 100 \qquad (1b)$$

$$\frac{C}{E_1+C} \times 100 \qquad (2a)$$

$$\frac{C}{E_2+C} \times 100 \qquad (3a)$$

$$\frac{2C}{E_1+E_2+2C} \times 100 \qquad (4b)$$

If it were desirable really to eliminate C from the denominator, this

could be done in some such form as:

$$\frac{C}{E_1+E_2} \times 100 \qquad (5)$$

but this is no longer a percentage with conveniently comparable values from 100 to 0, and it has no compensating advantages.

Finally, some such formula as this could be used:

$$\frac{E_1}{E_1+C} \times 100 = \frac{E_1}{N_1} \times 100 \qquad (6)$$

This is a percentage, but it is merely 100 minus index (2) and has the disadvantage of rating identity as 0 and complete difference as 100. There are, of course, comparable analogues for indices (1), (3), and (4).

TAXONOMIC LEVELS

An essential factor in all the indices of taxonomic resemblance is the hierarchic level of the taxa counted. It need hardly be emphasized that values of an index are comparable only if they are at the same taxonomic level and that the values will generally be higher the higher the level. Choice of too high a level for a given group of comparisons will give many or all values = 100, and too low a level will give many or all values = 0. It is almost always possible to find a level with most or all values less than 100 and more than 0, thus permitting meaningful comparisons. Determination of such a level is in itself a general indication of the degree of phylogenetic affinity involved in the comparisons.

As has been emphasized especially by Burt (1958), direct comparisons assume that the taxa really are comparable in the two faunas, that, for instance, genera are not notably more split in one than in the other. When taxa are evidently more split in one, that must be taken into account in interpreting the meanings of the indices. It may be added that index (2) has the additional advantage of tending to minimize effects of disproportionate splitting in the larger fauna.

Burt also finds it misleading that E (in symbols here used) includes forms that have relatives (at a higher taxonomic level) in both faunas as well as those that do not. For instance in comparing the Recent North American with the Asiatic mammalian fauna, *Odocoileus* has fairly close Asiatic relatives, *Antilocapra* does not, but both are counted in E. He proposes to include *Odocoileus* (and numerous other genera with Asiatic relatives) in C. If this adjustment is clearly understood and can be made consistently, it may indeed be useful. It should, however, be noted that the same purpose can be achieved more objectively by simply including analysis at an appropriate higher taxonomic level. In all of Burt's examples his objections would have been met more or less adequately by also giving indices at the subfamily or family level.

Another possible adjustment, still closer to Burt's purpose although he does not suggest it, would be to redefine C and E:

C_1', Number of genera (or other appropriate specified lower taxa)

in the first fauna (or sample) belonging to families (or other appropriate higher taxa) common to both.

C_2', Same for second fauna or sample.

E_1', Number of genera (or as specified) in the first in families (etc.) absent in the other.

E_2', Same for the second fauna or sample.

With these definitions, C', unlike C, is not a single number the same for both faunas, but indices comparable to those already discussed can readily be devised, e. g.:

$$\frac{C_1'+C_2'}{2N_1+2N_2-C_1'-C_2'} \times 100 \qquad (7)$$

$$\frac{C_1'}{N_1} \times 100 \qquad (8)$$

Index (7) is analogous (but not directly comparable) to (1), and (8) to (2). The different form of the denominator of (7) is necessary to keep this index analogous with (1) bu substituting the mean of C_1' and C_2' for C. A more logical but less convenient equivalent expression is:

$$\frac{(C_1'+C_2')/2}{N_1+N_2-[(C_1'+C_2')/2]} \times 100 \qquad (7a)$$

All the taxonomic indices, (1)—(8), are based on existing classifications. They necessarily reflect such subjectivity and lack of consistency as may affect those classifications. They also necessarily neglect relationships of possible importance for zoogeography (e. g. probable direction of spread between two areas) that are not involved in or expressed by classification. Like any other statistics, they cannot automatically solve any problem. They can order and quantify the data that enter into them and thus they assist in sound comparisons. After being calculated, they must still be interpreted in the light of knowledge of exactly what it is that each of them measures and in accordance with zoological and stratigraphic principles.

MEASURES INVOLVING ABUNDANCE OF TAXA

The indices hitherto mentioned, which include those in most general use, directly measure resemblance in taxonomy only. They take no overt account of the abundance of individuals in the taxa or of various other factors. In another sense, it is evident that faunas will be more similar if the taxa in common are relatively abundant in individuals and less similar if the most abundant taxa are those peculiar to each. *Ecological* differences between *taxonomically* similar faunas may, indeed, be largely reflected in the relative abundance of their taxa. For discussion of this point an additional set of symbols is needed.

I_{E1}, I_{E2}, Number of individuals of taxa not in common in first and second samples, respectively.

I_{C1}, I_{C2}, Number of individuals of taxa in common in first and second samples, respectively.

$I_1 = I_{E1} + I_{C1}$, $I_2 = I_{E2} + I_{C2}$, Total number of individuals in first and second samples, respectively.

$I_C = I_{C1} + I_{C2}$, Total number of individuals in the taxa in common.

The new and interesting idea of establishing an index in which individual abundance figures is due to Crusafont and Truyols (1958). As a preliminary trial they proposed the following index:

$$\frac{I_C : 2C}{I_1 : N_1 + I_2 : N_2} \times 100 \qquad (9)$$

The following algebraically identical formula may be somewhat easier to calculate and is more obviously indicative of the mathematical behavior of the index:

$$\frac{I_C N_1 N_2}{2C\,(I_1 N_2 + I_2 N_1)} \times 100 \qquad (9a)$$

In the original publication Crusafont and Truyols noted some disadvantages of this formula, for instance that it is not scaled from 0 to 100 (not being a percentage) and may give values far above 100, with a seemingly erratic scale making comparisons uncertain. In further illustration of its peculiarity, it suffices to point out that for two identical samples the value of this index is 50 while for nonidentical samples the value may also be 50 or higher or lower. In view of this and other inconvenient properties of the index (some noted in the original publication and some in subsequent unpublished discussion) it is preferable to seek another form of index based on the same general principle.

Among possible alternatives is:

$$\frac{I_C}{I_1 + I_2} \times 100 \qquad (10)$$

— the percentage of individuals in taxa in common among total individuals in both samples. This has a simple and rational scale, from 100 for complete identity to 0 for no resemblance, but it still has the drawback that differences in sizes of unbiased samples may result in different values of the index. That consideration suggests the use of:

$$\frac{1}{2}\left(\frac{I_{C1}}{I_1} + \frac{I_{C2}}{I_2}\right) \times 100 = \left(\frac{I_{C1}}{I_1} + \frac{I_{C2}}{I_2}\right) \times 50 \qquad (11)$$

— which is the mean of the percentages of individuals in common taxa figured separately for each sample. This also has the logical percentage scale from 100 to 0 and (if sampling is unbiased) it is not biased by sample size. This simple index thus has none of the serious weaknesses of (9) and (10) and is, I think, the best to use when an index of this particular kind is wanted. It should, of course, still be borne in mind that different indices measure *different* things and that a value of, say, 50 for (11) does not necessarily (although it may) indicate the same degree of resemblance as one of 50 for any other index.

Indices (9), (10), and (11) all have in common another weakness: they will not reliably reflect differences in relative abundance of taxa in faunas that are closely similar taxonomically. If, for instance, two samples have all their taxa in common, (10) and (11) will always be 100 and (9) will always be 50 even if the faunas are ecologically different and have different dominant taxa. But as Crusafont and Truyols point out, it is exactly this situation that calls for considering relative abundance and not only taxonomic resemblance. If taxonomic resemblance is moderate or low, indices (9)—(11) will not be likely to give any reliable information not inherent in the purely taxonomic indices. Thus (9)—(11) are least indicative when most needed.

Of several possible solutions to this special problem, perhaps the simplest is to use rank correlation of the taxa in common. The taxa are listed with the abundance or frequency of each in the available samples. For each sample each taxon is assigned a rank, 1 for the most abundant, 2 for the next, etc. For each taxon, the rank for one sample (it does not matter which) is then subtracted from the rank of the other, giving a difference, d. The rank correlation coefficient is then:

$$r_r = 1 - \frac{\Sigma (d^2)}{C (C^2 - 1)} \qquad (12)$$

This coefficient has a logical but somewhat peculiar scale, and its significance in this usage is not the same as in usual statistical correlation. Values range from +1 to —1. The value +1 reflects maximal resemblance, —1 minimal resemblance, and 0 indicates not absence of resemblance but an intermediate value. In order to make this formally similar to most of the previously discussed indices, with scales of 100 to 0, it may be transformed to:

$$50 + 50 r_r \qquad (13)$$

It must, however, again be remarked that a value for (13) does not indicate the same degree and sort of resemblance as the same value for any other index.

The coefficient (12) and index (13) leave entirely out of account the taxa that are not common to the two samples. These measures are therefore indicative only when most taxa are in fact common to the two. When that is not true, (11) may be used, or a purely taxonomic index, preferably (2), may be sufficient.

Measures (9)—(13) are all influenced both by purely taxonomic resemblance and by numbers of individuals in the common or noncommon taxa. Although (12) and (13) do not vary with numbers of taxa in either or both categories, they can only be used when in fact C is large, so that there is still a strong influence of numbers of common taxa. These are all compromise measures which either apply only under certain conditions as to taxonomic resemblance or seek to evaluate taxonomic and some sort of ecological resemblances simultaneously. They may correspondingly be ambiguous or difficult to interpret. They are, in particular, quite limited as measures of ecological resemblance, which frequently is largely or even almost completely indepen-

dent of taxonomic resemblance. None of these measures takes any account whatever of nontaxonomic ecological similaries or differences. Again, it is essential to know exactly what an index is actually measuring and to select an index that really is indicative for the particular problem in hand.

An important further point is that relative abundance of individuals in different taxa may in fact strongly influence the values of the apparently purely taxonomic indices (1)—(8). This is not the case if both faunas are completely known taxonomically and the question of sampling probability does not arise. If, however, either fauna is incompletely known, then the probability that a given taxon will appear in the corresponding sample will tend (other things being equal) to depend on the relative abundance of individuals in that taxon. Under these circumstances if the taxa in common are also more abundant, the taxonomic indices will tend to be larger than if these taxa were relatively rare. Thus in sampling situations such as are so common, especially with paleontological materials, all the indices, including those that are overtly taxonomic only, are likely also to be covertly affected by resemblances in individual abundances of taxa. This is not necessarily a disadvantage and may even be an advantage, but it is another factor to be taken into account in interpretation of the data.

Note on Sampling Effects

In the remarks just made, and also earlier in the paper, sampling effects have been mentioned in quite general terms. For fully reliable comparisons statistical confidence intervals should, of course, be established for any index used. For any of these indices, that is an exceptionally difficult problem that has not yet been worked out. The present notes are notes only, and not even by intention the last word on the problem. In the meantime use of the recommended indices, notably (2), is pragmatically justified because on the whole and with suitable caution they do give fairly consistent and meaningful results. To give them up and to rely on purely subjective comparisons until a complete statistical theory is available would be unjustified. The indices are objective data even when uncertainties remain in their statistical interpretation.

This problem does not arise for indices (1)—(8) when the faunas are completely known taxonomically. The population relevant here is a real and finite entity, quite distinct from any theoretical population of statistics. If the finite population, a list of taxa, is known, that is a direct and basic observation not affected by any considerations of probability. Then a value of, say, 78 for index (2) means exactly that. It is a measurement, not an estimate, of the true value and there is no probability at all that the true value is 79, 77, or any other figure than the one observed. If in a second such comparison a value 80 is obtained, then the difference between that and the first value of 78 is, so to speak, infinitely significant. There is no chance at all that the population values are the same. What this small difference means in terms of a particular problem is another matter; the difference, as such, exists.

When not only numbers of taxa but also numbers of individuals are taken directly into account [indices (9)—(13)], there is always a question of sta-

tistical estimation of probability. It is not likely ever to happen that *all* individuals in the populations can be counted, and there is bound to be some sample variance in proportions of the various taxa. The approach here is not too difficult, at least in theory, although the calculations would often be onerous. For any one taxon, confidence intervals for the relative frequencies of its individuals can be obtained by well-known statistical methods (e.g. Simpson, Roe, and Lewontin, 1960). Approximate confidence intervals for r_r [(12)] are also given in readily available texts.

The really serious problem is that of incomplete sampling of taxa, which directly or indirectly affects all the indices here mentioned. When all taxa are known, the confidence interval at any level is zero, as already mentioned. The intervals become larger, and apparently in a complex way, as the *proportion* of known taxa becomes smaller. But in this case the proportion itself is unknown, because the absolute number of taxa in the population is unknown. As already noted, the confidence intervals in terms of taxa are also indirectly influenced by the abundance of individuals and therefore by confidence intervals for relative abundance, which again greatly complicates the problem. There are, moreover, many other factors that may affect the probability of finding any particular taxon.

Finally, it is certainly true of some of these measures and perhaps of all of them that they are biased when based on an incomplete roster of taxa. That is, repeated samples of less than the full number of taxa will not tend to vary about, or to give as an average, the true value of the index in the population. Generally they will tend to give smaller values. A precise expression for this tendency would, again, be very complicated and has not been worked out.

All of this calls for further and better study. Nevertheless comparisons by the same index and with samples including about the same proportion of the populations' taxa are reasonably indicative. Indices based on a large proportion of population taxa also approach the population value nearly enough for most purposes.

AN EXAMPLE

Table 1 gives data for fairly good, representative fossil faunules from the same area, the relationships of which are resonably well understood. Table 2 applies most of the suggested measures to comparisons of these faunules. The results exemplify in some detail the characteristics of these indices already discussed in a more theoretical way.

If differences in facies are minimized in favor of purely taxonomic or phylogenetic resemblances, comparison (a-b) should give a high value, (b-c) also high, (b-d) considerably lower, and (c-d) somewhat lower still. Index (2) most nearly agrees with these expectations, although the discrepancy between (b-d) and (c-d) is perhaps greater than might have been anticipated. This may be related to the probability that both c and d include only small fractions of the taxa of the original populations so that all the indices for these two comparisons are relatively less reliable.

TABLE 1

Comparative data for genera of fossil mammals from different localities and horizons in the Paleocene of the Crazy Mountain Field, Montana. (Based on Simpson, 1937).

Datum (defined in text)	Faunal Comparisons			
	a-b	b-c	b-d	c-d
E_1	3	2	8	11
E_2	16	21	29	13
N_1	21	15	13	13
N_2	34	34	34	15
C	18	13	5	2
E'_1	0	0	1	6
E'_2	7	13	12	3
C'_1	21	13	12	7
C'_2	27	21	22	12
I_{E1}	3	7	43	104
I_{E2}	83	156	349	59
I_{C1}	55	72	50	3
I_{C2}	299	226	33	20
I_1	58	79	93	107
I_2	382	382	382	79
I_C	354	298	83	23

Samples from: a, Silberling quarry. b, Gidley quarry. c, Lebo surface localities. d, Scarritt quarry.

Comparisons: a-b, samples of unequal size similar in facies and age.
b-c, samples of unequal size and different facies but about the same age.
b-d, samples of unequal size and different (successive) ages with some resemblance in facies.
c-d, samples fairly close in size but distinctly different in both age and facies.

TABLE 2

Measures of faunal resemblance from the data of table 1

	Measures (as in text)	Comparisons (defined in table 1)			
		a-b	b-c	b-d	c-d
(1a)	$\frac{C}{N_1 + N_2 - C} \times 100$	49	36	12	8
(2)	$\frac{C}{N_1} \times 100$	86	87	38	15
(3)	$\frac{C}{N_2} \times 100$	53	38	15	13
(4)	$\frac{2C}{N_1 + N_2} \times 100$	65	53	21	14
(7)	$\frac{C'_1 + C'_2}{2N_1 + 2N_2 - C'_1 - C'_2} \times 100$	77	53	57	51
(8)	$\frac{C'_1}{N_1} \times 100$	100	87	92	54
(9a)	$\frac{I_C N_1 N_2}{2C\ (I_1 N_2 + I_2 N_1)} \times 100$	70	69	34	43
(10)	$\frac{I_C}{I_1 + I_2} \times 100$	80	65	17	12
(11)	$\left(\frac{I_{C1}}{I_1} + \frac{I_{C2}}{I_2}\right) \times 50$	85	75	31	14
(13)	$50 + 50r_s$	86	54	—[a]	—[a]

[a] For these two comparisons C is too small to permit meaningful use of this index.

For combined measures of taxonomic and ecological resemblances, (a-b) should be markedly high, and the other measures successively distinctly lower. Index (9) has this tendency but is still inadequate, while (10), (11), and, as far as it goes, (13) give reasonable results on this basis.

The whole table strikingly demonstrates the fact that comparisons cannot usefully be made from one index to another, although (3) does follow the same trends as (1). Indices (2), (3), and (4) tend to take the same value when N_1 is nearly equal to N_2, as in the comparison (c-d); (7) and (8) at quite a different level also tend toward the same value under this condition. Discrepancies in sample size strongly affect (1), (3), (4), and (7) but have little influence on (2) and (8).

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CRYSTAL CHEMISTRY OF β-SPODUMENE SOLID SOLUTIONS ON THE JOIN $Li_2O.Al_2O_3$–SiO_2*

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ABSTRACT. β-spodumene ($LiAlSi_2O_6$) is tetragonal, space group $P4_32_1$, $a = 7.5332 \pm 0.0008$ Å and $c = 9.1540 \pm 0.0008$ Å. It is isostructural with keatite (SiO_2) and forms an extensive solid solution towards SiO_2 and $LiAlSiO_4$. β-spodumenes of various compositions were synthesized and their unit cell edges determined.

On the assumption that β-spodumene is a "stuffed silica structure" based on the known keatite structure, probable locations of the introduced lithium atoms are established from geometrical considerations. From the change of unit cell volume with composition, it is deduced that, up to the spodumene composition, lithium atoms are inserted in cavities on the twofold axes; beyond this composition, additional cations are inserted in cavities in general positions in the structure. Channels extend throughout the structure connecting these cavities, providing paths along which lithium atoms can move under potential gradients. These channels account for the cation-exchange properties of β-spodumene.

Data are presented on the compositions of co-existing β-spodumene solid solution and β-eucryptite solid solution at 1350°C and 1 atmosphere and on the composition of β-spodumene solid solution co-existing with tridymite at 1350°C and 1 atmosphere.

INTRODUCTION

Spodumene, $LiAlSi_2O_6$, inverts to a high temperature modification on heating to sufficiently high temperatures. Many workers have noted that spodumene undergoes an irreversible physical change on heating, but it remained for Hatch (1943) to define this change correctly as a polymorphic inversion. Roy, Roy, and Osborn (1950) reported the low to high spodumene inversion occurred at 500°C and 10,000 psi H_2O pressure. The equilibrium inversion temperature at 1 atmosphere pressure, however, has not been satisfactorily defined because of the sluggish nature of the change at temperatures below 800°C without suitable fluxes. The slope of the inversion curve in terms of temperature and pressure is not yet known.

Phase equilibrium studies in the system Li_2O–Al_2O_3–SiO_2 have become increasingly important in recent years with the development of lithium-bearing ceramics, and since 1949 many significant papers have appeared on different aspects of the system. Throughout these studies some confusion has attended the exact nature of the high temperature modification of spodumene and the apparently wide compositional variation observed for the phase.

Current terminology designates the mineral spodumene as the low or α-form and the high temperature modification of $LiAlSi_2O_6$ as high or β-spodumene.

PHASE RELATIONS

The natural anhydrous lithium aluminum silicates all fall on the join $Li_2O.Al_2O_3$–SiO_2. Hatch (1943) and Roy, et al. (1950) have demonstrated that the phase relations along this join are essentially binary.

The original liquidus and solidus curves determined by Hatch (1943) have been slightly modified by Roy et al. (1950) and the present concept of the sub-solidus relations of β-spodumene along the join $Li_2O.Al_2O_3$–SiO_2 is essentially that published by the latter workers. Hatch (1943) demonstrated a

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large stability field for β-spodumene extending towards SiO_2 (β-spodumene–SiO_2 solid solution) and also towards $LiAlSiO_4$ (eucryptite) composition.

The present study, conducted at 1350°, has demonstrated that the stability field of β-spodumene solid solution is terminated towards SiO_2 by the two-phase field, β-spodumene solid solution plus tridymite, and is terminated towards $LiAlSiO_4$ composition by the two-phase field, β-spodumene solid solution plus β-eucryptite solid solution. This is in agreement with the findings of Hatch (1943) and of Roy, et al. (1950), although the field terminations found here do not agree exactly with either of the previous studies. Roy, et al. indicated that a narrow two-phase field β-spodumene solid solution plus β-eucryptite solid solution was a logical deduction but also noted: "That a gap exists has not yet been proved experimentally." We believe we have proved this point, and can now report the dimensions of the gap at 1350°C.

CRYSTALLOGRAPHY

Previous workers correctly recognized that β-spodumene crystals were tetragonal in their gross morphology. Keat (1954) pointed out the similarity between the x-ray powder diffraction data for β-spodumene and for the tetragonal silica polymorph, keatite.

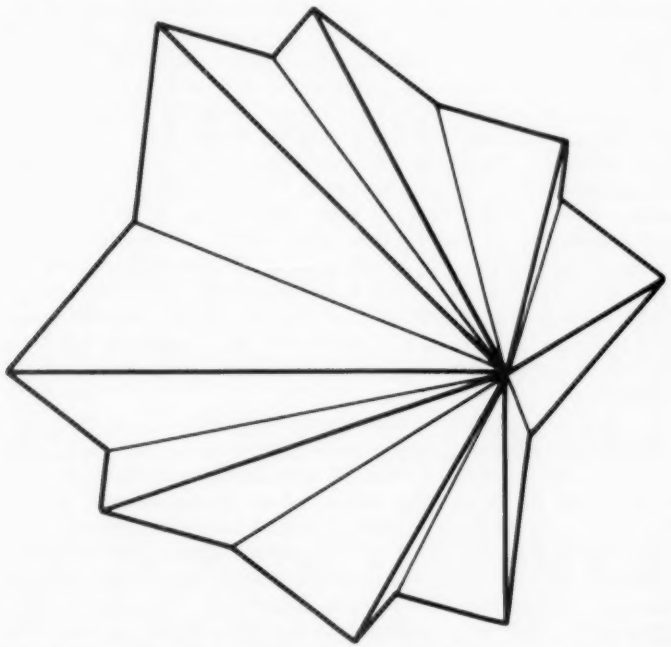

Fig. 1. Interpenetration twinned habit of crystals of β-spodumene solid solution of approximate composition $LiAlSi_4O_{10}$. Form $d\{011\}$; twin plane (111).

During our study of the phase relations in systems involving spodumene, small single crystals of β-spodumene and β-spodumene solid solutions were prepared. Some of these were found suitable for x-ray single crystal measurement. For this purpose crystals grown at 730°C and 2000 bars H_2O pressure for 48 hours, having an approximate composition of $LiAlSi_4O_{10}$, were selected. The crystals show the single form $d\{011\}$, producing an octahedral habit (for d, $\rho = 50°\ 18'$). They almost invariably occur as interpenetration twins on (111), so that they commonly have a peculiar rosette shape (fig. 1). By means of the Buerger precession method a tetragonal unit cell was found with $a = 7.50$ Å and $c = 9.03$ Å ($c:a = 1.204:1$) in the space group $P4_32_1$ (D_4^8) or its enantiomorph $P4_12_1$ (D_4^4). These data show a close relationship to those found in a recent single crystal study of keatite by Shropshire, Keat and Vaughan (1959), for which a unit cell was found with $a = 7.46$ Å and $c = 8.61$ Å, in the same space group. A comparison of the intensities of the *hk0* and *hhl* reflections for β-spodumene recorded on our Buerger precession photographs with those reported by Shropshire et al. for keatite shows the structures of the two crystals to be closely similar. Keat's (1954) suggestion that the two are structurally related is thus fully confirmed.

The unit cell of β-spodumene contains 4 formula units of $LiAlSi_2O_6$ (corresponding to 12 SiO_2 formula units for keatite). The unit cell dimensions for β-spodumene with stoichiometric composition were found by precision powder diffractometer measurements to be $a = 7.5332 \pm 0.0008$ Å and $c = 9.1540 \pm 0.0008$ Å at 25°C. These cell dimensions give a calculated specific gravity of 2.38, in good agreement with the measured value of 2.35 (Hummel, 1950).

Using the crystallographic data obtained from the single crystal x-ray studies, the x-ray powder diffraction data for β-spodumene were completely indexed (table 1). Several previous attempts had been made to index the powder pattern of β-spodumene. Roy, et al. (1950) suggested a tetragonal unit cell with $a = 6.31$ Å and $c = 8.32$ Å, but admitted that it did not give a satisfactory indexing of the powder data. Keat (1954) successfully indexed the β-spodumene pattern by analogy with keatite. He obtained a tetragonal unit cell with $a = 7.5$ Å and $c = 9.0$ Å, in general agreement with the present study. Plyushchev, et al. (1959) indexed β-spodumene on a tetragonal cell with $a = 13.15$ kX and $c = 11.64$ kX, but this cell has no apparent relation to the true cell. Previous measurements are gathered in table 1 for comparison with the present data.

β-SPODUMENE SOLID SOLUTIONS

Glasses of various compositions were prepared along the join $Li_2O.Al_2O_3$–SiO_2 in the same manner and from the same lithium carbonate, alumina and silica glass used by Stewart (in press). 50 to 100 mg charges, wrapped in platinum foil, were placed in a platinum wound muffle furnace from 1.75 to 3 hours at 1350°C. The glasses crystallized entirely within a few minutes, but it was observed that the first crystallization product was always a hexagonal β-eucryptite solid solution (or silica-O, Roy, 1959). The β-eucryptite solid solution rapidly inverted to a β-spodumene solid solution, so that within 15 minutes a homogeneous β-spodumene solid solution resulted.[1]

[1] This phenomenon has been studied in more detail in this laboratory by W. C. Phinney.

TABLE 1

X-ray powder diffraction measurements of β-spodumene and β-spodumene solid solutions. d(calc) computed for a unit cell having $a = 7.5332$ Å and $c = 9.1540$ Å. Intensities (I) by visual estimation from powder patterns.

			d (observed), Å					
I	hkl	d(calc) Å	$LiAlSi_2O_6$	75.21 wt% SiO_2 24.79 wt% $Li_2O.Al_2O_3$	Petalite, West. Aust. 1350°C, 1.75 hrs.	$LiAlSi_2O_6$ Roy et al. (1950)	Spodumene, 1100°C, 6 hrs. Plyushchev et al. (1959)	Keatite (SiO_2) Keat (1954)
	010	7.534	—	—	—	—	—	7.46
3	101	5.817	5.821	5.787	5.777	—	5.83	5.64
0.5	110	5.327	5.340	5.320	5.307	—	5.21	5.28
5	111	4.604	4.610	4.583	4.571	4.61	4.61	4.50
		—	—	—	—	—	4.35	—
6	102	3.912	3.920	3.881	3.868	3.97	3.89	3.72
	200	3.767	—	—	—	—	—	—
10	201	3.483	3.487	3.464	3.461	3.47	3.49	3.42
1	112	3.472	—	—	3.434	—	—	3.33
0.5	210	3.369	—	—	3.355	—	3.35	—
5	211	3.162	3.167	3.110	3.141	3.162	3.16	3.11
	202	2.908	—	—	—	—	2.93	—
1	103	2.828	2.858	2.796	2.793	2.927	2.85	—
1	212	2.713	—	2.705	2.695	—	2.72	—
0.5	220	2.664	—	—	2.656	—	—	—
3	113	2.648	2.628	2.628	2.617	2.609	2.65	2.516
	221	2.557	—	—	—	—	—	—
	030	2.511	—	—	—	—	—	—

TABLE 1 (Continued)

			d (observed), Å					
I	*hkl*	*d*(calc) Å	$LiAlSi_2O_6$	75.21 wt% SiO_2 24.79 wt% $Li_2O.Al_2O_3$	Petalite, West. Aust. 1350°C, 1.75 hrs.	$LiAlSi_2O_6$ Roy et al. (1950)	Spodumene, 1100°C, 6 hrs. Plyushchev et al. (1959)	Keatite (SiO_2) Keat (1954)
0.5	301	2.422	—	—	2.403	—	—	—
	310	2.382	—	—	—	—	—	—
	203	2.371	—	—	—	—	—	—
	311	2.306	—	—	—	—	—	—
4	222	2.302	2.304	2.294	2.287	2.327	2.30	2.246
1	004	2.289	2.291	—	2.258	—	2.27	2.148
4	213	2.262	2.264	2.248	2.238	2.261	2.23	2.174
	302	2.202	—	—	—	—	—	—
2	104	2.190	2.191	2.171	2.161	—	2.19	—
4	312	2.113	2.113	2.097	2.100	2.087	2.11	2.067
1	114	2.103	2.106	—	2.075	—	2.07	1.988
	320	2.089	—	—	—	—	—	—
	321	2.037	—	—	—	—	—	—
0.5	223	2.007	—	—	1.988	—	—	—
	204	1.956	—	—	—	—	—	—
5	303	1.939	1.938	1.928	1.922	1.936	1.94	1.879
1	322	1.901	1.899	—	1.888	—	—	—
	214	1.893	—	—	—	—	—	—
6	400	1.883	1.882	1.873	1.874	1.880	1.88	1.864
6	313	1.878	1.873	—	1.863	—	—	—
	401	1.845	—	—	—	—	—	—
0.5	410	1.827	—	1.818	1.817	—	1.79	—
	411	1.792	—	—	—	—	—	—

TABLE 1 (Continued)

			d (observed), Å					
I	*hkl*	*d*(calc)Å	$LiAlSi_2O_6$	75.21 wt% SiO_2 24.79 wt% $Li_2O.Al_2O_3$	Petalite, West. Aust. 1350°C, 1.75 hrs.	$LiAlSi_2O_6$ Roy et al. (1950)	Spodumene, 1100°C, 6 hrs. Plyushchev et al. (1959)	Keatite (SiO_2) Keat (1954)
0.5	105	1.779	—	—	1.757	—	—	—
	330	1.776	—	—	—	—	—	—
	331	1.743	—	—	—	—	—	—
1	402	1.742	1.742	1.735	1.731	1.738	1.74	—
	224	1.736	—	—	—	—	—	—
	115	1.731	—	—	—	—	—	—
0.5	323	1.724	1.723	1.714	1.710	—	—	—
2	412	1.697	1.697	1.689	1.686	1.697	1.70	—
	304	1.691	—	—	—	—	—	—
	420	1.685	—	—	—	—	—	—
2	421	1.657	1.656	1.651	1.648	1.673	1.65	—
	332	1.655	—	—	—	—	—	—
	314	1.650	—	—	—	—	—	—
4	205	1.647	1.646	1.634	1.627	1.631	1.63	—
1	215	1.609	1.608	—	1.590	—	—	—
	403	1.603	—	—	—	—	—	—
	422	1.581	—	—	—	—	—	—
	413	1.568	—	—	—	—	—	—
2	324	1.543	1.543	1.532	1.527	1.540	1.55	—
2	333	1.535	1.535	—	—	1.528	—	—
	225	1.509	—	—	—	—	—	—
	430	1.507	—	—	—	—	—	—
	050	1.507	—	—	—	—	—	—
0.5	106	1.495	1.494	1.482	1.476	1.505	1.50	—

The quenched charges were examined optically, but the very small grain size of the β-spodumene solid solutions precluded accurate optical measurements, other than a mean refractive index. Adequate optical measurements for β-spodumenes were presented by Hatch (1943). With larger crystals, produced in hydrothermal runs, it was possible to confirm observations by Hatch (1943) and others that β-spodumene is uniaxial positive. The optically negative β-spodumene mentioned by Roy, et al. (1950) is a β-eucryptite solid solution which forms metastably in the β-spodumene stability field as noted above.

All charges were also examined by x-ray powder diffraction photography, as this method was found much more sensitive in detecting small amounts of other phases in addition to β-spodumene, than examination by x-ray diffractometer. Once the homogeneity of a sample was established, from the optical and x-ray examinations, the unit cell edges of the β-spodumene solid solutions were determined by measuring the positions of the 400, 004 and 303 diffraction lines on an x-ray diffractometer, using NaCl[2] as an internal standard. All measurements were made with nickel filtered copper radiation. The results of these measurements are presented in table 2 and graphically in fig. 2.

TABLE 2

Unit cell edges of β-spodumene and β-spodumene solid solutions. Composition expressed as weight percent SiO_2, in the series $Li_xAl_xSi_{1-x}O_2$.

Composition weight percent SiO_2	x	Temp. °C.	Time (hours)	a, Å. ± 0.0008 Å	c, Å ± 0.0008 Å	d_{400} Å ± 0.0003 Å
62.04	0.3580	1380	3	7.5584	9.1720	1.9442
62.46	0.3539	1380	3	7.5548	9.1680	1.9430
62.89	0.3497	1350	1.75	7.5500	9.1656	1.9419
63.32	0.3455	1380	3	7.5452	9.1620	1.9411
63.73	0.3415	1350	1.75	7.5426	9.1608	1.9403
64.16	0.3373	1380	3	7.5396	9.1564	1.9392
64.58	0.3333	1350	1.75	7.5332	9.1540	1.9376
(spodumene composition)						
68.12	0.2990	1350	1.75	7.5224	9.1188	1.9345
71.66	0.2649	1350	1.75	7.5120	9.0964	1.9313
75.21	0.2310	1350	1.75	7.5020	9.0568	1.9267
78.75	0.1973	1350	1.75	7.4932	9.0348	1.9236
82.29	0.1639	1350	1.75	7.4832	9.0004	1.9179
Petalite	0.2000	1350	1.75	7.4976	9.0372	1.9240
Spodumene	0.3333	1350	1.75	7.5368	9.1696	1.9403

Petalite from Western Australia and spodumene from Newry, Maine, were also converted to β-spodumene by heating at 1350°C for 1.75 hours. The unit cell edges of the β-spodumenes so formed are presented in table 2. Neither the petalite nor spodumene have been chemically analyzed. If we assume them to have ideal formulas ($LiAlSi_4O_{10}$ for petalite and $LiAlSi_2O_6$ for spodumene) and plot the unit cell edges on figure 2, we obtain good agreement for the β-

[2] A unit cell edge of 5.64028 Å at 20°C was used for NaCl (Frondel, 1955).

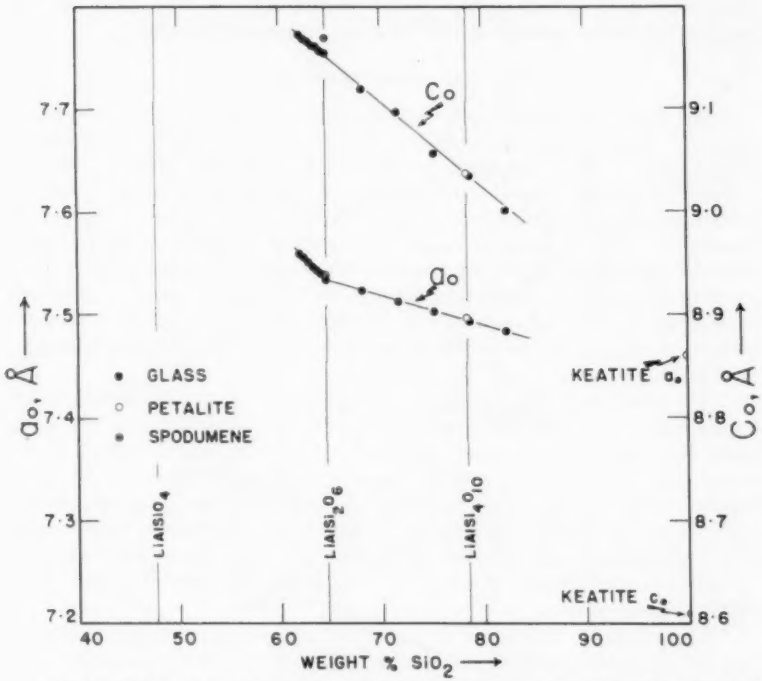

Fig. 2. Relationship between unit cell edges and composition of β-spodumene solid solutions. Prepared from synthetic glasses, petalite from Western Australia and spodumene from Newry, Maine. Values for keatite from Shropshire, et al. (1959).

spodumene solid solution formed from the petalite but unsatisfactory agreement for the Newry, Maine material. This suggests that the petalite is, in fact, close to the ideal formula, but that the Newry spodumene either deviates from the ideal in respect to its SiO_2 content or very possibly contains other alkali metal atoms replacing lithium in the structure, causing a consequent expansion of the unit cell.

Carefully prepared x-ray powder diffraction photographs demonstrated that no symmetry changes occurred throughout the β-spodumene solid solution stability field. We can state with certainty that β-spodumene and β-spodumene solid solutions have the same space group ($P4_32_1$) and are isostructural with keatite throughout the entire stability field at 1350°C. There are continuous changes in the unit cell edges of β-spodumene solid solutions with composition (fig. 2; table 2). Although we have not been able to make β-spodumene solid solutions more siliceous than 84.3 weight percent SiO_2, a solid solution series expressed generally by the formulation:

$$Li_xAl_xSi_{1-x}O_2$$

may be postulated ranging from $x = 0$ (keatite) to $x = 1/3$ (β-spodumene),

those solid solutions between 84.3 weight percent SiO_2 ($x = 0.1451$) and 100 percent SiO_2 being unstable at 1350°C and 1 atmosphere pressure.

CRYSTAL CHEMISTRY OF β-SPODUMENE

We have confirmed Keat's (1954) suggestion that β-spodumene has a structure closely analogous to that of keatite. This fact immediately classes β-spodumene with the so-called "stuffed silica structures" of M. J. Buerger (1954). From this viewpoint, β-spodumene may be regarded as being derived from the framework of the silica polymorph keatite by replacing a portion of the silicon atoms by aluminum atoms and "stuffing" an equal number of lithium ions into interstices in the framework to maintain charge balance. Fortunately, a complete structure determination of keatite has recently been carried out by Shropshire, Keat and Vaughan (1959). It is of interest to inquire as to what locations in this structure the introduced lithium and aluminum atoms occupy.

The structure of keatite, as illustrated in figure 3, consists of a network arrangement of twelve SiO_4 tetrahedra in the tetragonal unit cell. Of these, eight form spiral chains around fourfold screw axes in the centers of the lateral faces of the unit cell, and the remaining four tetrahedra lie on horizontal diagonal twofold rotary axes, cross-linking the spiral chains. Crystallographically, the lithium atoms (which in any case may occupy at random only a part of a symmetrically equivalent set of points) may go into the structure either in an eightfold general postion, or in a fourfold special position on the twofold rotary axes.

That the latter case is the primary one is demonstrated by the change in unit cell dimensions with lithium content. The plots of a and c versus composition (fig. 2) show surprising but internally consistent results. The a plot has a sharp break precisely at spodumene composition, and is linear on either side of the break; c on the other hand, does not break at this composition, and within the limits of the data is linear from 62 to 82.3 weight percent SiO_2. If the c versus composition plot does in fact break slightly at spodumene composition, greater precision of measurement will be required to demonstrate the point. In figure 2, straight lines have been fitted by eye to the points. These plots, together with a plot of d_{303} against composition (table 2) were used to determine the composition of β-spodumene solid solutions in mixtures.

It is apparent from these data that the mechanism of insertion of lithium atoms into the β-spodumene solid solution undergoes an abrupt change at the spodumene composition. At this point, exactly four lithium atoms are present in the tetragonal unit cell. The most reasonable explanation for the observed behavior is that at this point the lithium atoms just fill a fourfold set of symmetry-equivalent points in the structure. We must therefore look for these lithium atoms on the twofold rotary axes. Additional lithium atoms will be introduced in some other location in the structure, either (also) on the twofold axes, or in an eightfold general position.

We have attempted to find a probable location for lithium atoms on the twofold axis from geometrical considerations without recourse to an experi-

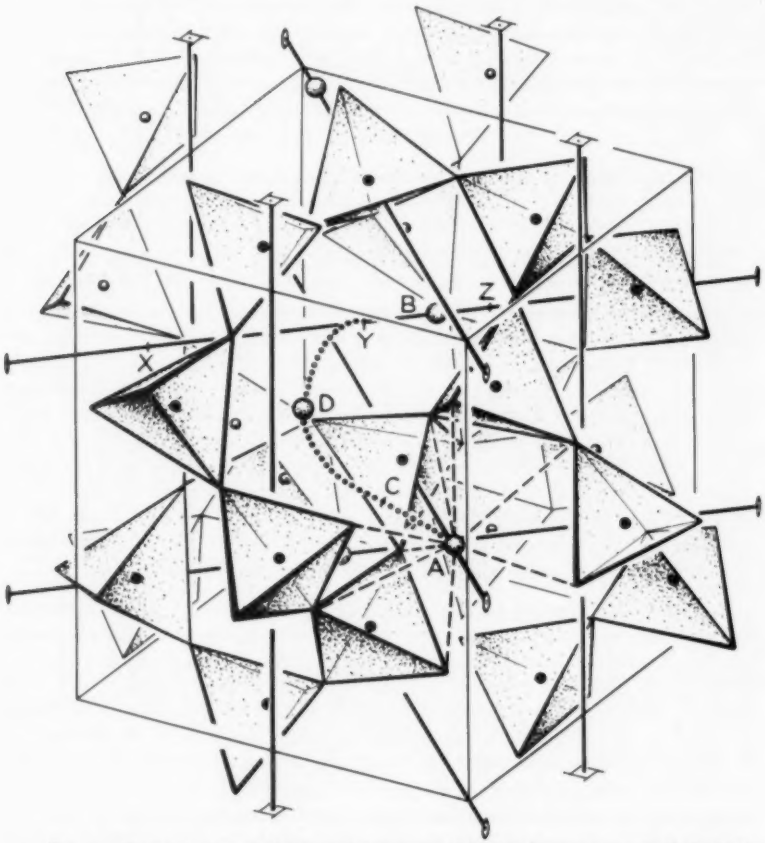

Fig. 3. Pictorial view of the crystal structure of keatite in space group $P4_32_12$. Probable lithium locations for β-spodumene solid solutions are indicated by shaded circles. Fourfold screw axes and twofold rotary axes are indicated, but twofold screw axes are not shown. Other features indicated by letters are described in the text. Origin at lower rear corner (hidden).

mental structure determination. Let us examine the twofold axis at $z = 3/4$ in the unit cell shown in figure 3. This axis is again shown in figure 4 in its repeat unit between the points 1, 0, 3/4 and 0, 1, 3/4 in the structure. The position of the silicon atom on this axis is shown at the right. The approach of adjacent oxygen atoms is shown by the hyperbolic curves drawn above the axis. At any point on this axis, a vertical line, by its intersection with these curves, shows the distance to the nearest oxygen atoms (in pairs). If the minimum interionic distance for lithium and oxygen is taken as 2.00 Å (as in Li_2O), it may be seen that the cation must lie somewhere between the points *X* and *Z*. In fact, it can only lie between *Y* and *B* to be in a reasonably uniform co-

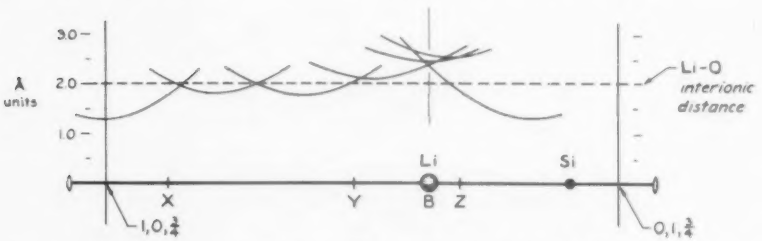

Fig. 4. The environment of the twofold rotary axis at $z = \frac{3}{4}$ in keatite, shown in fig. 3. Hyperbolic curves indicate approach of oxygen atoms to the axis (see text).

ordination without crowding. The largest cavity is found at the site *B* which is surrounded by eight oxygen atoms at distances of 2.34 (4), 2.41 (2) and 2.53 Å (2). Eightfold coordination is abnormal for lithium, which is commonly found in sixfold or fourfold coordination with oxygen, but in β-spodumene slight adjustments of the silica framework from the keatite arrangement probably occur which tend to reduce the eightfold coordination to six. The center of the cavity corresponds to position 4*a* in space group $P4_32_1$ with the parameter $x = 0.13$ (from the definition of the structure as given by Shropshire et al. 1959). We propose that lithium occupies this site at random in β-spodumene solid solutions (for example, sites *A* and *B*, fig. 3 and 4) up to $LiAlSi_2O_6$ composition for which these sites are completely filled.

In considering the possible location of lithium atoms in excess of four in the unit cell, we find that there are other cavities in the keatite structure. In fact, it is possible to trace a channel through the structure along a sharply curved path from site *A* on one twofold axis to its equivalent site *B* on an adjacent twofold axis. No point on this path, indicated by a train of small circles in figure 3, is approached by oxygen atoms closer than 1.95 Å. As pointed out by Zoltai and Buerger (1959), the structure of keatite can be viewed as an interlocking system of fivefold rings of SiO_4 tetrahedra. The site *A* faces two of these rings, and the channels, starting from this point, pass through their centers, one downward and one upward. On the latter path, the point *C* is thus at the center of a nearly regular pentagon of oxygen atoms at distances 2.07, 2.41, 1.99, 2.23 and 2.13 Å. At point *D* there is a cavity, roughly tetrahedral in shape, where the closest oxygen distance is 2.10 Å. It is quite possible that in the β-spodumene solid solutions richer in lithium than the spodumene composition, the lithium atoms are introduced at random into positions equivalent to site *D*. This is an eightfold general position 8*b* in space group $P4_32_1$ with parameters $x = 0.82$, $y = 0.49$ and $z = 0.68$. The insertion of lithium atoms into these positions, which are much smaller cavities than those on the twofold axis, accounts for the sudden enhancement of the rate of increase of the cell volume with lithium content at spodumene composition.

The channel originating at *A* reaches the adjacent twofold axis at point *Y* which is within the cavity associated with site *B* (equivalent to *A*). Thus it is apparent that the channels extend continuously throughout the structure.

This feature of the β-spodumene structure provides a considerable degree of freedom of movement for the lithium atoms under the influence of potential gradients. It is not possible to predict how the lithium atoms in the large cavities on the twofold axes will behave at higher temperatures, but their movement in this environment may be connected with the unusual negative thermal expansion which is characteristic of β-spodumene.

The chemical behavior of β-spodumene is of particular interest in relation to what we know of its crystal chemistry. One method of extracting lithium from natural spodumene involves its conversion to β-spodumene by heating and subsequent chemical leaching of lithium from the β-spodumene structure. White and McVay (1958) state that

> a definite change in crystal structure takes place during leaching of the β-spodumene but, by treating with LiOH under proper conditions, the leached β-spodumene can be reconverted, partially, to a form very similar to the β-spodumene.

Although we have not conducted experiments bearing on the point, we suggest that the reason lithium can be easily leached from and replaced in the β-spodumene structure lies in the existence of the channels described above, through which the cations can freely move. Thus, β-spodumene is endowed with cation exchange properties (presumably restricted to Li^+ and H^+ ions) analogous to those of the zeolites. α-spodumene, on the other hand, has a pyroxene-like structure which is much tighter (compare the specific gravities of α-spodumene, 3.2, and β-spodumene, 2.38) and has no channels through which the cations may escape. Natural spodumene is therefore not affected by acid solutions.

As far as the aluminum atoms are concerned, there is no way of predicting whether they will tend to replace silicon preferentially in the sites on the twofold axes or in the general positions. Owing to the high temperatures associated with the formation of β-spodumene, they probably actually substitute at random over all silicon positions.

The structural hypotheses proposed above may eventually be substantiated by crystal structure analysis, at least as far as locating the lithium atom on the twofold axis is concerned. Owing to the low scattering power of lithium to x-rays, three-dimensional methods would probably be required, presuming that suitable untwinned crystals of β-spodumene can be prepared for the diffraction intensity measurements.

DATA IN THE TWO PHASE FIELD

Runs less siliceous than 62 weight percent SiO_2 crystallize as a mixture of β-spodumene solid solution and β-eucryptite solid solution at 1350°C. A run containing 60.76 weight percent SiO_2 was crystallized for 3 hours at 1350°C and the composition of the resulting β-spodumene solid solution determined by x-ray analysis. The β-spodumene solid solution had a composition of 62.3 weight percent SiO_2 ($a = 7.5576$ Å, $c = 9.1720$ Å, $d_{303} = 1.9438$ Å). The coexisting β-eucryptite solid solution had a composition of 59.8 weight percent

SiO_2 (a = 5.2340 Å and d_{212} = 1.6346 Å), as determined from the unpublished data of Skinner on the change in unit cell edge versus composition of β-eucryptite solid solutions. From 59.8 weight percent SiO_2 to at least 47.69 weight percent SiO_2 ($LiAlSiO_4$ composition) β-eucryptite solid solution is the sole stable phase at 1350°C.

Runs more siliceous than 84.3 weight percent SiO_2 crystallize to β-spodumene solid solution plus tridymite at 1350°C. A glass containing 85.83 weight percent SiO_2 crystallized to tridymite plus a β-spodumene solid solution with an averaged composition of 84.3 weight percent SiO_2 (a = 7.4788 Å, d_{303} = 1.9176 Å, c not determined) and a glass containing 89.37 weight percent SiO_2 crystallized to tridymite plus a β-spodumene solid solution containing 84.2 weight percent SiO_2 (a = 7.4796 Å, c and d_{303} not determined). Longer runs at the same temperatures did not change either the compositions or relative amounts of the phases present. Thus, at 1350°, the β-spodumene solid solution stability field falls between 62.3 weight percent SiO_2 and 84.3 weight percent SiO_2. It is hoped that further work will define the stability field of β-spodumene solid solution and β-eucrypite solid solution at lower temperatures.

ACKNOWLEDGMENTS

We wish to acknowledge, with thanks, the assistance and many helpful discussions we have had with our colleagues David B. Stewart and William C. Phinney.

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DISTRIBUTION OF THE FRESHWATER CLAM *PISIDIUM ULTRAMONTANUM*; A ZOOGEOGRAPHIC INQUIRY*

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ABSTRACT. *Pisidium ultramontanum* Prime, a freshwater clam of the Sphaeriidae, is a relict species found living only in northeastern California and south-central Oregon. Pliocene and Pleistocene fossils, however, document its former occurrence in the Snake River drainage as far east as southeastern Idaho. Analysis of the distribution of other mollusks and fishes associated with *P. ultramontanum* reveals a common pattern, which can be traced by faunal links from one drainage basin to another. These links form a chain which leads from Walker Lake in western Nevada across Eagle Lake and the upper Pit River, California, to Klamath Lake, Oregon; thence across Fossil Lake and the Malheur basin, Oregon, to the Snake River; and through Gentile Valley and Bear Lake, Idaho, to Utah Lake, Utah. All of the relict mollusks and fishes of lakes in the northern Great Basin, Snake River drainage, and Klamath River drainage are related to this pattern of distribution. These lakes are therefore thought to have been connected, although not necessarily all at once.

INTRODUCTION

The small freshwater clam *Pisidium ultramontanum* Prime (pl. 1) has the most restricted distribution of any member of the family Sphaeriidae in North America. It is known living only in northeastern California and south-central Oregon, but recent work by the U. S. Geological Survey has shown that it occurred much farther east during the Pliocene and Pleistocene, as far as southeastern Idaho (fig. 1). These fossil occurrences, some as old as early Pliocene, demonstrate that the present local occurrence of *Pisidium ultramontanum* is not due to relatively recent origin, but that it is a relict species.

Faunal ties between Pliocene and Pleistocene assemblages of the Snake River valley, southern Idaho, and the Recent fauna of northeastern California and southern Oregon are evident among freshwater snails in addition to this *Pisidium*. The correlation between these occurrences and those of other mollusks and fishes shows the significance of their pattern of distribution. The former drainage connections inferred from this pattern help to explain other geologic and zoogeographic data which have not been related previously.

This paper has benefited from criticism by S. Stillman Berry, Redlands, California and P. E. Cloud, Jr., U. S. Geological Survey; and from discussions of fish distribution with Carter R. Gilbert, U. S. National Museum. Léon Croizat, Caracas, Venezuela, contributed essential parts of the method of analyzing distribution through his monumental "Panbiogeography" (Croizat, 1958). H. B. Herrington, Westbrook, Ontario, identified much of the *Pisidium* material.

OCCURRENCE OF PISIDIUM ULTRAMONTANUM

Recent and fossil occurrences of *Pisidium ultramontanum* (Prime, 1865) are shown in figure 1, and the precise localities listed at the end of this paper. The distribution pattern is considered significant, and not at all fortuitous, for several reasons. If *P. ultramontanum* had been subject to casual dispersal, one would expect that it would be more widely and erratically distributed, especial-

* Publication authorized by the Director, U. S. Geological Survey.

ly since it is known to be at least as old as early Pliocene. The fossil shells are all from near large streams of the present day, and the three known Recent localities are in lakes or medium-sized streams. Other mollusks and fishes have distribution patterns correlated with that of *P. ultramontanum*, implying some historical significance.

The zoogeographic and geologic evidence pertinent to the distribution of *Pisidium ultramonanum* will be considered first in regard to the conspicuous Idaho-California discontinuity, and then to more local details.

HYDROGRAPHIC LINKS BETWEEN IDAHO AND CALIFORNIA

Former river and lake connections between southern Idaho and northeastern California are suggested by the distribution of *Pisidium ultramontanum* (fig. 1). Supporting evidence consists of other faunal ties between these two areas, of geologic evidence that the present course of the Snake River is relatively young, and of zoogeographic and geologic evidence that the Snake formerly flowed to the Pacific Ocean independent of both the Columbia and Sacramento rivers.

Faunal ties.—Two fossil freshwater snails associated in southern Idaho with *Pisidium ultramontanum* share similar affinities with northeastern California and vicinity. *Ceriphasia acutifilosa* (Stearns, 1890) of the prosobranch family Pleuroceridae, lives in the upper Pit River and in Honey Lake basin, northeastern California (fig. 1). An undescribed close relative has been found

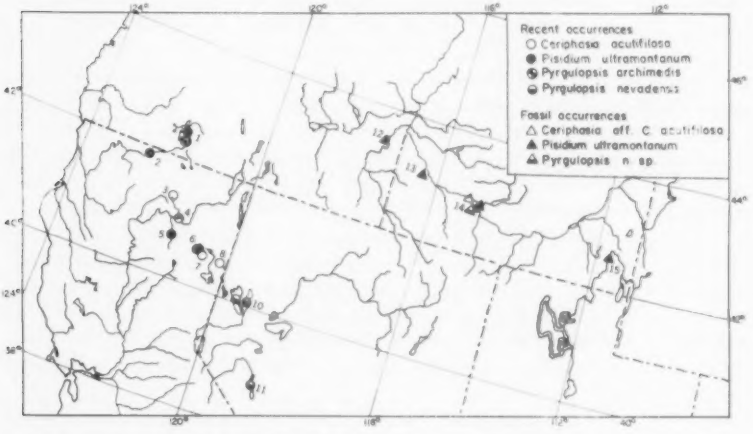

Fig. 1. Distribution of *Pisidium ultramontanum*, *Pyrgulopsis nevadensis* group, *Ceriphasia acutifilosa*, and *C.* n. sp. Sources of information are U. S. Geological collections and references cited in text. Numbered localities are as follows: 1, Klamath Lake. 2, Hornbrook. 3, Headwaters of Fall River. 4, Fall River Mills. 5, Hat Creek (formerly Canoe Creek). 6, Eagle Lake. 7, Head of Willow Creek. 8, Secret Valley. 9, Pyramid Lake. 10, Winnemucca Lake. 11, Walker Lake. 12, T. 23 S., R. 43 E., Malheur Co., Oregon. 13, T. 5 S., R. 1 E., Owyhee Co., Idaho. 14, Hagerman-Glenns Ferry area, Elmore and Twin Falls counties, Idaho. 15, Gentile Valley.

in late Pliocene and early Pleistocene deposits, along with *Pisidium ultramontanum*, in southern Idaho. As emphasized by Goodrich (1944, p. 7-9), members of this family move only through their aquatic habitat. The only other member of this species group (Goodrich, 1942) is *C. occata* (Hinds), of the Sacramento River drainage in California; the relations of the Idaho fossil shells are thus entirely to the southwest.

The freshwater snail *Pyrgulopsis* (Prosobranchia, Hydrobiidae) is represented by a number of species in North America, of which three are closely related. These are *P. archimedis* Berry (Berry, 1947) of Klamath Lake, Oregon; *P. nevadensis* (Stearns), living in Pyramid Lake, Nevada, and subfossil in Walker and Winnemucca lakes, Nevada (Berry, 1947; Call, 1884); and an undescribed late Pliocene form from southern Idaho (Taylor, in preparation). Distribution of these species is shown in figure 1. As with the Pleuroceridae, this family is restricted to perennial water bodies, and no evidence is known to suggest casual dispersal.

Recency of present Snake River course.—Both geologic and biogeographic data suggest that the present course of the Snake River through Hell's Canyon to the Columbia River is relatively young. Barbed tributaries, different degrees of maturity of the Snake River canyon, and differences in the canyon gradient indicate that relatively recent stream capture created the present canyon (Wheeler and Cook, 1954). Living mammals on either side of Hell's Canyon are but slightly different, whereas those of southwestern Idaho and southeastern Oregon are more distinct (Cook and Larrison, 1954).

Independent course of Snake river.—The former Snake River was thought by Wheeler and Cook (1954, p. 534) to have flowed through northeastern California into the Feather River, and thence into the Great Valley. Comparison of late Pliocene molluscan faunas from Idaho and California suggests instead that the Snake flowed into the Pacific independent of the Sacramento system. Mollusks from the basal part of the Tulare formation of the Kettleman Hills (Pilsbry, 1935; Woodring and others, 1940), from the Santa Clara formation of northwestern California (University of California and U. S. Geological Survey collections), and from the Tehama formation of northern California (University of California collections) are closely related. Late Pliocene mollusks from southwestern Idaho are so distinct from the Central Valley assemblages as to imply no direct aquatic connection at this time. Distribution patterns of living mollusks and fishes suggest closer relationships with the Klamath drainage.

DISTRIBUTION PATTERNS FROM SOUTHERN OREGON TO WESTERN NEVADA

In the broad outlines of distribution showing faunal affinities between southern Idaho and northeastern California, *Pisidium ultramontanum* has been shown to be similar to *Pyrgulopsis* and *Ceriphasia acutifilosa*. More local details of distribution in the area where these three are found from southern Oregon to western Nevada (fig. 1) will now be considered. The distribution patterns are most conveniently analyzed according to the gaps between pairs of

now separate drainage basins. Faunal ties between the Klamath River and upper Pit River, and upper Pit River and western Great Basin are suggested by the distribution of other animals beside the three mollusks mentioned already.

Klamath River-Upper Pit River.—Both fishes and aquatic mollusks show close relationships between the Klamath basin and upper Pit River. Fishes common to these two drainages are *Entosphenus*, a dwarf non-parasitic lamprey known only in these two regions; stream forms of the chub *Siphateles bicolor;* and the sculpin *Cottus klamathensis*. *Cottus tenuis* of the Klamath Lake basin is closely allied to *C. asperrimus* of the upper Pit River (Robins and Miller, 1957).

"That none of the endemic Sacramento forms reported from the upper Pit drainage (e.g., *Hesperoleucus*, *Mylopharodon*, *Ptychocheilus*, and *Hysterocarpus*) ocurs in the Klamath suggests that the connection between these two basins was severed prior to the invasion of the upper Pit region (through headward erosion by Pit River) by these Sacramento types" (Robins and Miller, 1957, p. 230).

Both *Pisidium ultramontanum* and *Ceriphasia acutifilosa* are found in the upper but not the lower part of the Pit River (fig. 1). *P. ultramontanum* in Klamath Lake and the upper Pit River has a distribution similar to that of the fishes, but *C. acutifilosa* shows Pit River-Great Basin relationships.

Upper Pit River-Great Basin.—Outside of the Klamath drainage, the only Recent occurrences of *Pisidium ultramontanum* are in the upper Pit River drainage and in Eagle Lake (fig. 1). This lake occupies an isolated basin between the upper Pit River and the Great Basin. It is more intimately related to the Lahontan part of the interior drainage than to the Pit, for according to Hubbs and Miller (1948, p. 38): "The fish evidence definitely calls for a surface-water connection between Eagle Lake and Lake Lahontan waters."

Occurrence of *Ceriphasia acutifilosa* in the upper Pit River and western Lahontan drainages (fig. 1) evidently represents a pattern like that of *Pisidium ultramontanum*, although different in detail. Fishes and these two mollusks thus overlap in range so as to link Klamath Lake, the upper Pit River, and the western Lahontan basin. The Recent occurrences of *Pyrgulopsis* (fig. 1) are thus related by the distributions of a variety of aquatic animals.

Klamath Lake-Great Basin.—Distribution of *Pyrgulopsis* and *Ceriphasia*, which are associated with *Pisidium ultramontanum* both living and fossil, has

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PLATE 1

Pisidium ultramontanum Prime. Link River, NW¼ sec. 30, T. 38 S., R. 9 E., Klamath Co., Oregon. D. W. Taylor, Oct. 4, 1959. Drawings by Mrs. Elinor Stromberg, U. S. Geological Survey.

Figs. 1-3. External views (dorsal, posteroventral, and lateral) of one specimen (USNM #143200). 4.65 mm long, 4.29mm high, 2.60 mm broad.

Figs. 4-5. Internal views (right and left valves) of another specimen (USNM #143115). 5.00 mm long, 4.72 mm high.

PLATE 1

Pisidium ultramontanum Prime

been shown to accord with fish distribution and geologic evidence. It is now appropriate to examine the occurrence of other fishes and mollusks to see if the same pattern recurs. Two lacustrine relicts, both occurring in Klamath Lake with *Pisidium ultramontanum* and both morphologically set off sharply from their relatives, are suggestive subjects for inquiry. These are the lake sucker, *Chasmistes*, and the freshwater snail *Carinifex*.

Chasmistes is represented in Klamath Lake by three described species. "These are comparable among living forms only with *Chasmistes liorus* of Utah Lake (Bonneville) and *C. cujus* of Pyramid Lake (Lahontan)" (Hubbs and Miller, 1948, p. 68). Fossil occurrences of the lake sucker consist of subrecent dried remains from Humboldt Cave, and of fossil bones from Fossil Lake and Lower Klamath Lake (Hubbs and Miller, 1948, p. 41, 68, 74). Fossil and Recent distribution is shown by figure 2. The Recent occurrence in

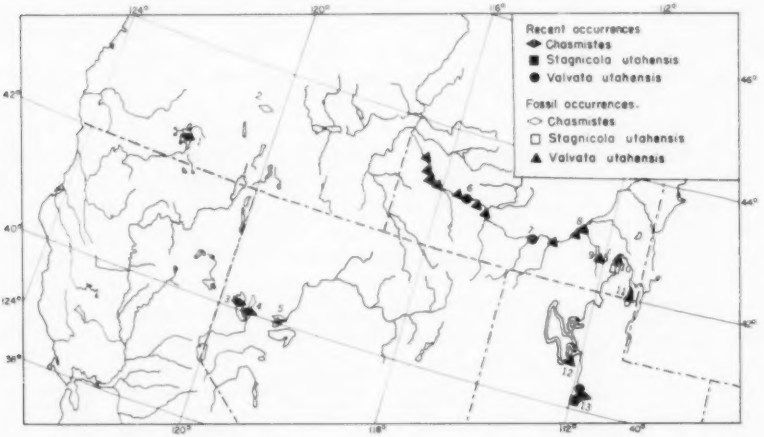

Fig. 2. Distribution of *Chasmistes*, *Stagnicola utahensis*, and *Valvata utahensis*. Sources of information are Hubbs and Miller (1948) and Snyder (1918) for *Chasmistes*; U. S. Geological Survey collections for *Stagnicola* and *Valvata*. Numbered localities are as follows: 1, Klamath Lake. 2, Fossil Lake. 3, Pyramid Lake. 4, Winnemucca Lake. 5, Humboldt Cave. 6, Snake River, NW¼ sec. 4, T. 6 S., R. 8 E., Owyhee Co., Idaho. 7, Snake River, NW corner sec. 22, T. 10 S., R. 23 E., Minidoka Co., Idaho. 8, American Falls Reservoir. 9, Marsh Creek valley. 10, Gentile Valley. 11, Bear Lake. 12, Saltair (from core). 13, Utah Lake.

Klamath and Pyramid Lakes is that of *Pyrgulopsis*, and links the Klamath and Lahontan basins presumably along the Pit River-Eagle Lake route suggested by *Pisidium ultramontanum*. The Utah Lake occurrence will be considered later, along with the eastern area of occurrence of *P. ultramontanum*. The fosil localities are nearby the Recent occurrences except for the one in Fossil Lake This will be discussed after the distribution of *Carinifex* (fig. 3).

Living representatives of *Carinifex* are included in only two species or species-groups. The monotypic *Carinifex minor* Cooper is restricted to Clear

Lake, California (Hanna, 1924). The varied *C. newberryi* group has a far wider distribution, occurring living or fossil over much of the western United States. The correlation with the distributions of *Chasmistes*, *Pyrgulopsis*, and *Pisidium ultramontanum* is striking (figs. 1-3), and shows again the Klamath Lake-Lahontan basin link by way of the upper Pit River and Eagle Lake. Like *Chasmistes*, *Carinifex* has an eastern area of distribution to be discussed subsequently, and a Fossil Lake occurrence treated now.

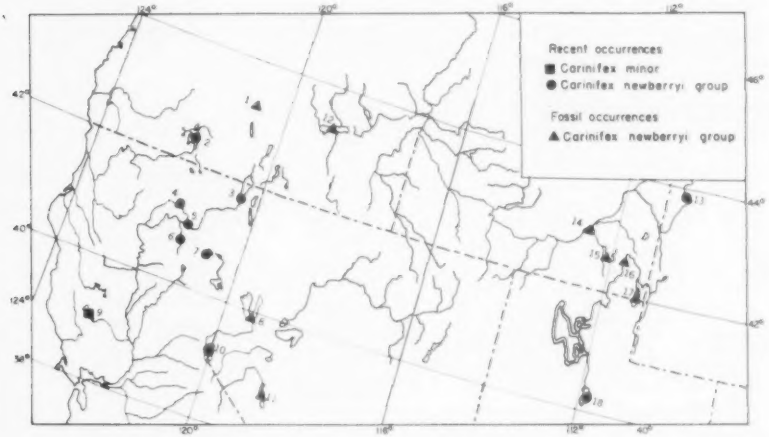

Fig. 3. Distribution of *Carinifex*. Sources of information are U. S. Geological Survey and U. S. National Museum collections, and references cited in text. A few fossil occurrences of the *C. newberryi* group in southeastern California (Baily and Baily, 1951-52) are not shown on the map. These occurrences are correlated with those of the minnow *Siphateles* (Hubbs and Miller, 1948, p. 79), which has faunal affinities with the Lahontan basin. Numbered localities are as follows: 1, Christmas Lake. 2, Klamath Lake. 3, Goose Lake. 4, Headwaters of Fall River. 5, Pit River. 6, Hat Creek (formerly Canoe Creek). 7, Eagle Lake. 8, South end of Winnemucca Lake. 9, Clear Lake. 10, Lake Tahoe. 11, Walker Lake. 12, Divide between Harney and Malheur lakes. 13, Jackson Lake. 14, American Falls Reservoir. 15, Marsh Creek Valley. 16, Gentile Valley. 17, Bear Lake. 18, Utah Lake.

Klamath Lake-Fossil Lake.—Fossil *Chasmistes* at Fossil Lake, Oregon (Hubbs and Miller, 1948, p. 74) and fossil *Carinifex newberryi* (Lea) at nearby Christmas Lake (Call, 1884, and U. S. Geological Survey collections) both suggest faunal affinity to Klamath Lake. The only living fish in the present drainage basin of these fossils which has zoogeographic value is a chub which seems most similar to a Klamath basin form (Hubbs and Miller, 1948, p. 74). A late Pleistocene pluvial lake may have drained into the Columbia River (Hubbs and Miller, 1948, p. 73), but the distribution of these animals gives no such indication.

DISTRIBUTION PATTERNS FROM EASTERN OREGON TO WYOMING AND UTAH

The eastern fossil occurrences of *Pisidium ultramontanum* in Oregon and Idaho (fig. 1) are now in separate drainages, those of the Snake River and

Bonneville basin. Local details of this pattern of distribution will be analyzed by pairs of drainage basins as before.

Malheur basin-Snake River.—Fossil *Carinifex* in the Malheur basin (fig. 3) forms a link between the Fossil Lake drainage to the west and Snake River drainage to the east. The modern Malheur basin fish fauna shows very close affinities with the Snake River drainage, in harmony with geologic evidence that the Malheur basin has been separated by lava flows from the Columbia basin rather recently (Hubbs and Miller, 1948, p. 75).

Snake River-Bonneville Basin.—Analysis of the faunal ties between the Snake River and Bonneville basin is more revealing when the Bonneville basin is separated into local areas. Utah Lake, Bear Lake, and a former Pleistocene lake in Gentile Valley all have endemic species as well as mutual affinities and ties with the Snake River drainage. These three lakes will be considred from south to north, beginning with Utah Lake.

Utah Lake.—*Pisidium ultramontanum* has not been found in Utah Lake, but *Chasmistes* and *Carinifex* both live there. These two are associated with *P. ultramontanum* in Klamath Lake, Oregon; and they both occur at Pyramid Lake, Nevada, along a chain of faunal links including *P. ultramontanum* and associates. In like manner other relicts in Utah Lake form links which connect with the occurrences of *P. ultramontanum* to the north.

Two snails described by Call (1884) from Utah Lake have been considered endemic there for a long time: *Valvata utahensis* and *Stagnicola utahensis.* The *Valvata* is now known to live in the Snake River, and its fossil occurrences (fig. 2) show that it was associated with *Pisidium ultramontanum* during the Pleistocene in southwestern as well as southeastern Idaho.

Stagnicola utahensis is still known living only in Utah Lake. The single fossil occurrence is of late Pleistocene age in Gentile Valley, Idaho, along with *Pisidium ultramontanum* (fig. 2).

Bear Lake.—Bear Lake, Utah-Idaho, is unusual in having fishes (three endemic) but no moluusks (Henderson, 1931). Abundant fossil shells around the shore of the lake (Henderson, 1931; Baily and Baily, 1951-52; U. S. Geological Survey collections) show affinities to those of Utah Lake and Pleistocene deposits in Gentile Valley.

The three endemic fishes are included in two genera of whitefish. *Prosopium spilonotus* (Snyder) is related to the widespread *P. williamsoni* (Girard) of the western United States; *P. spilonotus* (Snyder) is rather isolated; and *Coregonus gemmifer* (Snyder) is the only species of the genus in the Pacific drainage (Moore, in Blair and others, 1957, p. 67-69).

The mollusks *Carinifex newberryi* and *Valvata utahensis* are part of the series of faunal ties between Utah Lake and the Snake River Valley (figs. 2-3). Two more species are endemic, but are most nearly related to other extinct forms from Gentile Valley, associated with *Pisidium ultramontanum.*

Gentile Valley.—Studies of the late Pleistocene history of Bear Lake and Gentile Valleys, southeastern Idaho, indicate that the Bear River formerly

flowed into the Snake River. Damming by lava flows formed a lake in Gentile Valley which spilled over the divide into the Bonneville basin. Later downcutting has drained the lake and established the Bear River in its present course (Bright, unpublished).[1] Mollusks from the deposits of this former lake include *Pisidium ultramontanum*, *Valvata utahensis*, *Stagnicola utahensis*, and *Carinifex newberryi*. The distributional affinities of these species are evident from figures 1-3. Other extinct genera and species are either endemic or related to Bear Lake forms.

BASIC UNITY OF THE DISTRIBUTION PATTERNS AND THEIR AGE

Inquiry into the Recent and fossil distribution of the freshwater clam *Pisidium ultramontanum* has revealed that similar patterns are repeated by freshwater snails and fishes. All of the relict mollusks and fishes of lakes in the northern Great Basin, Snake River drainage, and Klamath River drainage are related to one pattern of distribution which can be traced by fanual links from one basin to another. These links form a chain which leads from Walker Lake in western Nevada across Eagle Lake and the upper Pit River, California, to Klamath Lake, Oregon; thence across Fossil Lake and the Malheur basin, Oregon, to the Snake River in southern Idaho and western Wyoming; and from the Snake River through Gentile Valley and Bear Lake, southeastern Idaho, to Utah Lake, Utah. All these areas are thought to have been connected by a series of rivers and lakes, although not necessarily all at once.

The age of this pattern of distribution cannot be determined precisely, but it is probably as old as Pliocene and at least as old as early Pleistocene. This estimate is based on the degree of differentiation of isolated populations and on the probable middle Pleistocene age of the Snake River capture. Stronger evidence comes from the evident lack of dispersal of many of the relict forms even when new and presumably suitable areas for existence became available.

The restricted distribution and lack of fossil record for *Carinifex* and *Pyrgulopsis* led Call (1884) to consider them recent immigrants into the Great Basin. In Utah *Carinifex* was found living in Utah Lake, but in none of the Lake Bonneville deposits. Call (1884, p. 20) concluded, "It is probably a comparatively recent addition to the fauna of the Bonneville Basin". *Pyrgulopsis nevadensis* (Stearns) lives in Pyramid Lake, and is found subfossil at Walker Lake. "Since careful collection has failed to reveal the form in Lahontan beds or Lahontan tufas, the hypothesis is justifiable that it has been introduced into each of these lakes since they became independent bodies" (Call, 1884, p. 27). *Valvata utahensis* and *Stagnicola utahensis* live in Utah Lake of the Bonneville basin, but likewise are not known from Lake Bonneville deposits. In the Lahontan basin *Carinifex* is known fossil, but only at the south end of Winnemucca Lake and at Walker Lake. All these snails are lacustrine, and one might think that they spread throughout Lakes Lahontan or Bonneville. Yet the record shows that they did not.

[1] Bright, R. C., Geology of the Cleveland area, southeastern Idaho; 1960 M.S. thesis, Univ. Utah.

Similar persistence and lack of spread is shown by mollusks and fishes in the Pit River and Klamath River. The lamprey *Entosphemus*, species of sculpin (*Cottus*), *Pisidium ultramontanum*, *Ceriphasia acutifilosa*, and *Carinifex* have all remained in the headwaters of these streams.

The distribution patterns and degree of local endemism of the fishes and mollusks discussed in this paper suggest that each species or species-group had a former distribution which included the Pleistocene and Recent occurrences. The patterns of these occurrences is inherited from an earlier pattern which antedates modern drainage basins and topography. Utah Lake, Pyramid Lake, Klamath Lake and others are local havens where an early Pleistocene or Pliocene habitat has persisted, while it was eradicated or strongly modified by block faulting, drainage shifts, desiccation and the like in intervening areas. Perhaps not all of the relicts are as old as *Pisidium ultramontanum*, which is known from the early Pliocene, but their common distribution pattern has modern antiquity.

LOCALITIES FOR PISIDIUM ULTRAMONTANUM (PRIME)

This list includes only localities represented by material examined by the writer or by H. B. Herrington.

Recent localities

Klamath Co., Oregon. NE¼ sec. 22, T. 36 S., R. 7 E. Upper Klamath Lake 0.6 mile south of Modoc Point. D. W. Taylor, Oct. 5, 1959.

Klamath Co., Oregon. NW¼ sec. 30, T. 38 S., R. 9 E. Link River (outlet of Upper Klamath Lake) at bridge. D. W. Taylor, Oct. 4, 1959.

Siskiyou Co., California. Klamath R., Hornbrook. W. O. Gregg, coll.

Shasta Co., California. Hat Creek (type locality). Prime (1865) gave the locality as "Canoe Creek", but according to Hanna (1924) the name is now Hat Creek. Types were credited to the Smithsonian Institution by Prime (1865), but as of December, 1959, these specimens could not be found. The only extant types are probably the cotypes in the Museum of Comparative Zoology, Harvard University (Herrington, 1950; Johnson, 1959).

Lassen Co., California. South end of Eagle Lake. D. W. Taylor and M. C. McKenna, Aug. 29, 1950.

Pleistocene localities

Franklin Co., Idaho. USGS Cenozoic loc. 21137. SE¼ sec. 7, T. 12 S., R. 41 E. ± 5400′ elev. R. C. Bright, 1957. Early Wisconsin age.

Franklin Co., Idaho. USGS Cenozoic loc. 21139. SE¼ sec. 7, T. 12 S., R. 41 E. ± 5150 elev. R. C. Bright, 1957. Early Wisconsin age.

Franklin Co., Idaho. USGS Cenozoic loc. 21141. NE¼ sec. 1, T. 12 S., R. 40 E. R. C. Bright, 1957. Early Wisconsin age.

Elmore Co., Idaho. USGS Cenozoic loc. 20103. NW¼ sec. 20, T. 6 S., R. 10 E. 3075′ elev. D. W. Taylor, 1956. Middle Pleistocene.

Pliocene localities

Elmore Co., Idaho. USGS Cenozoic loc. 20125. NW¼ sec. 14, T. 5 S., R. 10 E. 2615′ elev. H. E. Malde and D. W. Taylor, 1956. Late Pliocene.

Elmore Co., Idaho. USGS Cenozoic loc. 19930. Center of sec. 32, T. 5 S., R. 10 E. 2550′ elev. D. W. Taylor, 1956, 1957. Late Pliocene.

Elmore Co., Idaho. USGS Cenozoic loc. 19931. NW¼ sec. 5, T. 5 S., R. 10 E. 2515′ elev. D. W. Taylor, 1956-1959. Late Pliocene.

Elmore Co., Idaho. USGS Cenozoic loc. 21531. NE¼ sec. 31, T. 7 S., R. 10 E. 3000′ elev. H. E. Malde, 1958. Late Pliocene.

Owyhee Co., Idaho. USGS Cenozoic loc. 19130. N½ sec. 19, T. 5 S., R. 1 E. D. W. Taylor, 1955. Late Pliocene.

Malheur Co., Oregon. USGS Cenozoic loc. 21175. NE¼ sec. 21, T. 23 S., R. 43 E.

3400′ elev. R. E. Corcoran, 1955. Early Pliocene.
Malheur Co., Oregon. USGS Cenozoic loc. 21332. NE¼ sec. 28, T. 23 S., R. 43 E.
3500′ elev. R. E. Corcoran, 1955. Early Pliocene.

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ORIGIN AND USE OF THE WORD "SHALE"*

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ABSTRACT. Shale is a word of Teutonic origin that developed its meaning of "laminated clayey rock" in the English mining district of Derbyshire. The first recorded use is in 1747, but clearly the term is much more ancient. By the first few years of the 19th century, it had superseded a number of local names as well as the classically derived schistus, which the early natural philosophers were accustomed to use. By the end of the 19th century, it was nearly the only term applied to clayey rocks of Jurassic age and older in England and to clayey rocks of all ages in the United States, except for the coastal plain strata of Cretaceous and Tertiary age. The word is now looked on with some disfavor because we use it both as the class name for fine-grained rocks and more precisely as the name for laminated clayey rocks. Both of these uses, however, seem permissible and justifiable.

The origin of the word "shale" is lost in the dusty beginnings of the mineral industry in England. Digging into the question exposes some of the roots of geologic terminology at the beginning of the 19th century and reveals layers of usage, recognition of which bears on problems of naming and classifying rocks. This brief review of the origin of the word "shale" and the development of its usage is based chiefly on literature available in the library of the U. S. Geological Survey. More reference material exists but has not been examined. This review, incomplete as it is, may remind us that words are a creation of man which, like children, are likely to turn out differently than expected.

The first explicit statement relating to the origin and geologic use of the word "shale" seems to be that of John Farey in 1811 (Derbyshire rept., v. 1, p. 443, cited by *Oxford English Dictionary*[1]):

> It is not uncommon with colliers to call any argillaceous stratum in very thin laminae by the name of shale.

The next explicit statement is that of Lyell (1833, App., p. 80):

> SHALE. A provincial term, adopted in geological science, to express an indurated slaty clay. *Etym.* German *schalen*, to peel, to split.

Subsequent compilers of glossaries such as Humble and Page seem to have followed Lyell on origin as well as meaning. Humble (1840, p. 233) gave *schale* as the source word; he perhaps lost the final *n* by error because he correctly gives *schale* as the source for shell. Page (1865, p. 405) refers to *schalen*. Gresley identifies the British regions that seem to be the sources for many terms used by miners, but he does not suggest either a source or origin for the word "shale" (1883, p. 216). Lyell's definition of shale and its etymology have been noted in the American literature by Twenhofel (1937, App. I., p. 93).

This etymology, however, is denied by the *Oxford English Dictionary*, which says:

> There is no sufficient reason for the common view that [the noun shale] is [from] the German [noun] *schale* . . . , which is not used in this sense, (the German equivalent being *schieferthon* 'slate-clay'): *schale* however occurs [in German] for a thin layer of ore or stone[2] . . .

Arkell and Tomkeieff (1953, p. 105) states only that the word "shale"

* Publication authorized by the Director, U. S. Geological Survey.

[1] Murray, J. A. H., Sir, ed., 1914, A new English dictionary on historical principles, v. 8, pt. 2, S-SH, by Henry Bradley, p. 606: Oxford, Clarendon Press, 10 v.

[2] This perhaps is the source of shell in the "sand and shells" of so many drillers logs.

> Probably [is] a derivative of the verb shale, obsolete except in [English] dialect, which means 'shell off,' 'decorticate.'

Under scallett (p. 102), they remark

> Presumably the word is a variety of scald, scalléd, meaning scabby, flaky, from the Old Teutonic root *skal*, from which shale and shill were derived.

Webster's New International Dictionary, second edition, gives the etymology as

> Middle English *shale*, *schale*, *schal* from Anglo-Saxon *scealu*, *sealu*, shell, husk, drinking cup, balance; akin to Old High German *scala* shell, husk, Middle High German *schal*, German *Schale*, Danish and Swedish *skal*, a shell.

The etymologists thus differ among themselves as to details but are agreed that the term has a Teutonic source and seem to imply that the term is an ancient one, perhaps brought into England during the Teutonic invasions, which began in the 5th century.

Lyell's mistake in deriving shale from the German verb *schalen* is inexplicable. He had visited Europe in 1828 and 1829 and was familiar with much German literature. In his glossary, Lyell refers only half a dozen or so words to German sources, among them blende, feldspar, and gneiss, as well as shale. These words seem to be peculiar choices because most of the early mineralogists were as diligent in citing the synonomies of mineral names (see Kirwan, 1784 or Jameson, 1816) as paleontologists are today. The abundant synonomies of Jameson would have made it clear to Lyell that slate-clay (1816, v. 1, p. 418) or bituminous shale (v. 1, p. 434) were varieties of *Schiefer* in the German of Werner. Aikin (1815, p. 242) should have been an acceptable source for this information, if Lyell's vigorous adoption of Hutton's views made him ignore the works of such ardent supporters of Werner as Kirwan and Jameson.

The linguistic source of the word "shale" may be left to the etymologists. It is now of interest to us only to examine some of the names used for clayey rocks in English before the word "shale" appears in the literature.

The first mention I have found of the rock we call shale is in the logs of holes drilled for coal in Yorkshire in 1639 by Thomas Waike. These logs were communicated to the Royal Society in 1699 by Dr. Martin Lister (1699 [1809], p. 354). The following extract is typical:

	Ft.	In.
A cowshot coloured[3] stone	6	0
Black metal	1	6
Cowshot coloured stone	3	9
Coal mixed with metal	0	9
A blue metal	1	6
Coal	2	9
Cowshot coloured stone	6	0

[3] From cowshuts, a stock dove, and therefore gray (Arkell and Tomkeieff, 1953, p. 31).

From its position in relation to the coal bed in Mr. Waike's log, metal clearly refers to shale. Waike also uses the words "ramel" and "slate" in logs of other holes. The *Oxford English Dictionary* cites J. Morton, (1712, Nat. Hist. Northampt., p. 129) as defining rammel [sic] as a "stone unfit for building

because it cleaves or shales into many small uneven pieces." Arkell and Tomkieff (1953, p. 110) find the word "slate" is derived from the Old French *esclat*, a splinter, and that it appeared in literature with the meaning roofing slates in the first half of the 14th century.

Shale associated with coal apparently was sometimes called slate, as it still is. In 1676, Hodgson ([1809], p. 359), in speaking of a fire in a coal mine near Newcastle says, "There is also a sort of mineral called slate, which is partly coal, partly alum-stone, partly marcasite . . . " Colwall says (1678 [1809], p. 458), "Alum is made of a stone . . . of a bluish colour, and [that] will cleave like Cornish-slate." The rock names used by both Hodgson and Colwall were not the native terminology for rocks and perhaps they were consciously using the terms of the educated. Marcasite, for instance, probably was obtained from Agricola (Bandy and Bandy, 1955, p. 17) or directly from classical sources. Brassil (Hooson, 1747, entry Brassil) was an ancient English name for iron sulfide. The ancient name for Cornish slate was killas (Pryce, 1778, p. 323), and the name still is used (Wells and Kirkaldy, 1951, p. 173).

Clay rocks were mentioned by Lister (1683) in his "ingenious proposal for a new sort of map," which seems to have been the first suggestion for the making of geologic maps and is a very clear statement of the uses to which they can be put. His proposal is accompanied by a table in which are described different kinds of clays. Under pure clays, "such as [are] soft like butter to the teeth, and [have] little or no grittishness . . . ", he included the "cow-shot" clay, or the "soap scale lying in coal mines" (1683 [1809], p. 85).

In the southwest of England, the word for the rocks directly over coal was clives (singular cliff). Strachey said (1719 [1809], p. 402), "These clives vary much in hardness . . . ; so likewise . . . in colour . . . ; wherever they are met with, coal is sure to be found under them, though not always worth the digging." In 1725, Strachey ([1809], p. 119) equates coal-clives [sic] with slate, a term that apparently was acquired during an examination of coal mines in Scotland and Northumberland.

Shale appears as a noun and receives its first definition in what seems to be the earliest English miner's guide. Published in 1747 by William Hooson, who characterizes himself as "a Derbyshire miner," the book explains "not only the terms used by miners, but also the theory and practice of [the] art of mining, more especially of lead-mines." The book was written in Hooson's old age (see epilogue following the entry "Yockings"), and he claimed to have more than 40 years experience in mining. The definition merits full quotation:

SHALE

> A tolerably hard and black substance, lying most commonly on some places next to the Lime-Rock, most of it will cut pritty well, rising in thin Pieces: the Joynts in it are of a red colour, or gankey, when it is sunk some space into it is mixt with binds, and when it is drawn up to the Day, it melts and crumbles into small, and into much lesser Pieces;* it is subject to Water, Veins will sometimes put up into it, and bear Ore, otherwise it seldom fails in any considerable Vein, but there will breath up a Spar leading in the form of a Ribb, and probably Smuts of Ore in it; this Shale seldome requires any Timber to support it, it is very strong and Sulpherous Work, and apt to breed the Ground-Damp, without a good Wind to throw it out.
>
> * As the Shale lies on a Level or Diping, so may be expected the Lime Rock to lie, that lies under it.

With respect to the origin of the word, this definition is like an iceberg: the most important part is out of sight. Hooson used the word in several other places in his book (see entry under Lidds, for instance), and it always is used as a common term and one that had long been known, needing no other explanation. Mawe (1802) did not include the word in his glossary of Derbyshire mining terms, but he used shale in the text, mostly apologetically, I infer, and usually as a synonym for schistus, a word that he clearly preferred even though shale might be first on his tongue (p. 12). Mawe also pointed out that the Derbyshire mining district had been active since the time of the Romans (p. 6). The antiquity of workings in the district is emphasized by the Hoovers (1950, footnote, p. 85), who pointed out that the first records of the district begin in 1288. They also found the mining customs of Derbyshire to be distinctly different from other mining districts in England and to be of undoubted Saxon origin. They believed that the ratification of these customs by the Normans caused the survival of one of the few Saxon institutions in England. Mawe's references to Saxon and German miners (p. A2, 46, and 109) seem to apply to the miners imported to England in the 16th and 17th centuries. A readily apparent Saxon survival is Hooson's use of the word "groove" for mine. By Mawe's time, groove had become grove (1802, p. 205). This is remarkably similar to the modern German *Grube*, which has the same meaning. Thomson (1814, p. 418) found groove used for mine in northern England and commented on its similarity to the Swedish word *grufve*, which he said "has the same meaning and . . . sound."

A tantalizing hint as to the possible antiquity of the word "shale" is contained in the rhymed account of "King Coal's Levee", the second edition of which was published in 1819. The title page reads, "King Coal's Levee, or Geological Etiquette, with explanatory notes. To which is added The Council of the Metals. by John Scafe, Esq." The following leaf reads "To the Geological Society of London this Humble Attempt to render more popular a science of which daily experience proves the increasing utility." The author acknowledges his indebtedness to Conybeare and to Buckland for technical assistance. John Scafe is not listed (Woodward, 1908, p. 268-285) as a member of the Geological Society, and I am not certain that he is the author of more than "The Council of Metals," which seems to be an addition made for the second edition. The verses describe a reception held by King Coal and Queen Pyrites. After the geologic formations of England are presented in stratigraphic order, beginning with Duke Granite and ending with Jack Clay and the retinue of Baron Basalt, Shale appears (p. 32):

> But from that group stept forward shiv'ring Shale,
> And grac'd his bow with compliments quite stale:
> Such as, mid hoops and ruffs, by every beau
> Were drawled around, two centuries ago.
> "Where has he liv'd?" Pyrites cried:—"*Unique.*"
> "Oh! quite a gem!—decidedly *antique!*"
> "An old acquaintance;" straight the King replied,
> And bade old Shale stand snugly by his side.

These lines suggest that their author believed the term "shale" to have become current in the early 1600s.

Shale, although of Teutonic origin, seems to have come from the early

Saxon invasions rather than being imported with Saxon miners. The earliest recorded entry of German miners into England is mentioned by the Hoovers (1950, footnote, p. 283) as having taken place in 1561 during the reign of Queen Elizabeth. Taylor (1799, p. 361) states that the technique of blasting was brought into English mining by Germans in about 1670. By at least 1556, Agricola was describing the Mannsfield copper slates as *Lapis aerosus fissilis*, which he translated into German as *schifer* [sic] (Hoover and Hoover, 1950, footnote, p. 127). It would seem that these miners would have brought with them the term "*Schiefer*," or cognates of it, rather than shale. Shale would have acquired its present meaning on English soil.

For reasons that are not evident, the word "shale" became the most widely accepted name for laminated clayey rocks. In arriving at this position, shale superseded a number of names, probably local both in origin and usage, that stemmed from the jargon of miners. Some of these names are listed below. A few names not strictly applicable to laminated clayey rocks are included simply because they are interesting.

Basses—Slaty clay with a bluish-black color and usually containing pyrite. May be coaly and grade into slaty coal or combined with petroleum and grade into cannel coal (Aikin, 1811, p. 197). Strachey (1719 [1809], p. 401) mentions the verb "to basset" for "to crop out" as having come from the north of England. This may be related to the fact that the hard shale above or below a coal bed is what is most often seen on outcrop, the softer coal being weathered back and covered.

Bat—Coal strata "have between them a bat, or bed, of a peculiar sort of matter, about the thickness of a crown piece." Below the coal strata are "different metalline strata; as a black substance called the Dun-row bat." The bats seem to take their names from the overlying coal beds (Chambers, 1751, article Coal). Shale is not mentioned.

Binder—A miner's term from the Pembrokeshire coal field translated by De La Beche simply as shale (1823, p. 19).

Binds—"Beds of Stone that always lie mixt with a certain thickness of Shale, or Clay between them; . . . they are commonly black Stone, very hard and sharp, the lesser or thinner the Shale, or Clay-Beds are between them, the more nearer they resemble the Nature of the approaching Ouges . . . " (Hooson, 1747). The Ouges are the firm hard rock that borders the veins, and in Derbyshire are mostly limestone.

Clod—Indurated clay; compact, dull, and smooth but somewhat meagre [harsh, rough?] to the touch (Aikin, 1811, p. 196).

Clunch—An indurated clay that is glossy, unctuous, tending to slaty texture (Aikin, 1811, p. 196). This clay sounds similar to the soap scale of Lister (1683 [1809], p. 85). Mawe (1802, p. 22) mentions a similar word, "clutch," as the colliers' term for the rock above coal.

Duns—A miner's term for a masive clay (Buckland and Conybeare, 1824, p. 253) found in southwest England. In this district, a fissile clay above coal is called clives, as has been mentioned. Buckland and Conybeare considered duns to be "obviously of the same origin as the German *Thon*." The miners in southwest England also used another term of possible German origin. Strachey

(1719 [1809], p. 401) mentions the phrase "appears to the day" as a miner's term. "To the day" was also used as the general directional term for the mouth of the mine or the surface of the ground in Derbyshire (Hooson, 1747, entry Day). The usages are remarkably similar to the modern German terms "*zu Tage ausgehen*" meaning crop out, and "*Tagebau*" for open-pit mine, as called to my attention by Curt Teichert (oral communication, November 1959). If duns comes from the German *Thon*, it does so by way of a German influence that is not readily apparent.

Dunstone—Translated by De La Beche simply as shale (1823, p. 19).

Griddle—"A thin hard bed of slate-clay or sandstone" (Thomson, 1814, p. 418).

Shiver—"Slate-clay approaching to a shale" (Thomson, 1814, p. 418). This word may have been adopted from German *Schiefer*, according to the *Oxford English Dictionary*, which cites the first usage of shiver from 1729. Farey (1814, p. 261) equates shiver and shale.

Slat—"Impressions of various kinds of plants [are found] particularly in the stratum of earthy sl[illegible], which always lies immediately on the coal stratum . . ." (da Costa, 1757 [1809], p. 123).

The general acceptance of the term "shale" by geologists presumably was sometime prior to 1814, because Thomson (1814, p. 418) defines the miner's specialized words "in the common mineralogical language which has been adopted in this country." In 1808, Farey was using shale as a common lithologic term for a clayey rock, and he also was using shale as a stratigraphic name in the combinations "coal-shale" and "limestone-shale." Farey hardly could have influenced English mineralogy very much because he was a practical mineral surveyor who was never accepted by the Geological Society of London. In fact, he was deliberately snubbed, it appears, and a paper that he offered to the Society in 1812 was treated brutally. (See Farey, 1813, footnote, p. 55, for a powerful expression of rightous indignation.) Perhaps Thomson had reference to an unrecorded action of a committee established by the Geological Society in 1808 to "remove the confusion that now prevails" in mineralogical and geological nomenclature (Woodward, 1908, p. 23). Woodward adds that no records appear to have been preserved of any decision reached by the committee.

In the late 18th and early 19th centuries, the term "schist" was widely used by the educated for rocks that we would call shale. Kirwan (1784, p. 86) uses schisti as a group term for "argillaceous fissile stones." Hutton (1788, p. 245) refers to the "bituminus schistus" in which ironstones occur, and schistus also is used by Playfair (1802, p. 12). This usage probably extends back to Pliny, who wrote of *lapis schistos* (*Oxford English Dictionary*), but it just as easily could have sprung independently from the knowledge of Latin and Greek possessed by the natural philosophers of the 18th century. The usage of the word does not come from Agricola, to whom schistos was an iron mineral (Bandy and Bandy, 1955, p. 8; Hoover and Hoover, 1950, p. 111).

Probably Lyell was largely responsible for the general acceptance of shale as a broad geological rather than a provincial mining term. His *Principles of Geology* went through at least 9 editions between 1830 and 1853, and his

Elements of Geology, and the succeeding *Manual*, went through at least 4 editions between 1838 and 1852. They were the most widely read geological texts of their time. In these books, shale is quite generally asociated with stratigraphic names, such as "Wenlock shales," and applied to nearly all English clayey rocks of Jurassic age or older; only the Cretaceous and Tertiary strata of England seem to lack shale. Lyell's books were printed in several American editions and he spent the years 1841 and 1842 in the United States. His influence in the United States was large.

It is not practicable to review in detail the development of terminology for fine-grained rocks in the United States, nor is it necessary, for the literature is readily available. The general trend prior to 1850 seems to have been to use shale for almost any clayey rock of Paleozoic age; afterwards the term came to be applied to many clayey rocks of all ages.

When Hayden and Meek began their explorations of the rocks of Cretaceous age in the Missouri River country, they were meticulous in calling the rocks known to us as the Pierre shale, and fine-grained rocks in younger strata, by the name "clay." (See Hall and Meek, 1855, p. 405.) Hayden is reasonably consistent in this usage in his later writings. Shale seems to have been attached to the Cretaceous rocks after the conflicts of interest between Hayden's Territorial Survey and King's Survey of the 40th parallel became apparent. King (1878, p. 298), in recognizing Hayden's priority in naming the Cretaceous rocks of the Western Interior region (Meek and Hayden, 1857, and 1862, p. 419), says that he appealed to Hayden "for a name to represent the three groups combined [Ft. Benton, Niobrara, and Ft. Pierre]. Accepting his suggestion of the name of Colorado . . . " On p. 306, King states that the Ft. Pierre group consists of grayish-black carbonaceous shales and marls, and nearly black arenaceous clays. In 1881, however, Stevenson writes (p. 95), "The Colorado group was proposed by Mr. King to include the Ft. Benton, Niobrara, and Ft. Pierre shales of Meek and Hayden." From this time on, nearly all stratigraphic units of mainly clay-sized rocks in the Western Interior region have been called shale, except for strata of Tertiary age, and even among these are found the Green River and Sentinel Butte shales.

Exposed rock units of Cretaceous and Tertiary ages in the southern and eastern coastal plain of the United States still are called clay, but many rock units of these ages on the Pacific coast are called shale. The expansion of the oil industry and rapid accumulation of subsurface data soon led to the general designation of fine-grained rocks in wells as shales. This was entirely natural because the exact nature of the rock could not easily be determined from cuttings, and may have been of no consequence. Electric logs give only an approximate identification of rock types.

J. Volney Lewis, in 1924, pointed out that the word "shale" was properly applied only to laminated clayey rocks (p. 568) and the validity of this usage was emphasized by Bradley (1931, p. 7). These suggestions came at a time when the study of sedimentary rocks was becoming more detailed and quantitative. The Journal of Sedimentary Petrology was founded in 1931. Since then, the nomenclature of sedimentary rocks has received much scrutiny.

This trend toward greater precision (and rigidity) in nomenclature tends

to blind us to an historical fact. Shale is the generally acepted name for the class of fine-grained rocks, of equal standing with sandstone and carbonate as names for major rock groups. We are committed, I think, to the continued use of the word "shale" for this purpose, and I see no reason to try to avoid it. Certainly our comprehension is broad enough to include two meanings of the word "shale": First, the reasonably precise meaning of "laminated clayey rock" to which the origin of the word entitles it, and second, the meaning of the "general class of fine-grained rocks," which our historical use of the word bequeaths to it.

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ON FISH REMAINS IN LACUSTRINE SEDIMENTS

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ABSTRACT. The uppermost 15 cm of the deep-water sediments of Little Round Lake, Ontario, contain an abundance of fish bones and scales mostly derived from *Perca flavescens*. The average numbers found per square meter of sediment surface were 16,000, 650, 400, and 205, for scales, vertebrae, arches (neural and hemal), and ribs, respectively. The deep-water sediments of two other lakes were devoid of fish remains. The number of scales found in the surface sediments of a fish hatchery pond was 6 percent of the number expected from fish that had died in the pond.

INTRODUCTION

Literally thousands of sediment cores have been collected from lakes and bogs, but only in one instance have these been reported to contain fish remains.[1] The one published case on record is that of Lagler and Vallentyne (1956) who found two isolated fish scales in 7,000 to 8,000 year old sediments from Linsley Pond, Connecticut. These scales were found by accident, and not as the result of premeditated search. From these single findings we suspected that fish remains might be more common in the deep water sediments of lakes than had hitherto been thought. The work here reported was undertaken to determine the abundance of fish remains in the deepest surface sediments of three lakes and one pond.

METHODS AND MATERIALS

The water-bodies selected for study were: Fayetteville Green Lake, located about 10 miles east of Syracuse, New York; Lake Opinicon, situated near the town of Chaffey's Lock in eastern Ontario; Little Round Lake, about 3 miles north of the town of Sharbot Lake in eastern Ontario; and Pond H, one of a series of artificial ponds at the Cornell Fish Hatchery, Fayetteville Green Lake is meromictic, and was examined precisely for that reason. It has a maximum depth of 59 m, with the level of the chemocline at 19-20 m (Eggleton 1931, 1956). Lake Opinicon is a shallow, holomictic, eutrophic lake with a maximum depth of 11 m (Curran, et al., 1947). The lake supports high populations of large-mouth bass, *Micropterus salmoides*, and the sunfish, *Lepomis macrochirus* and *L. gibbosus*. Serious limnological studies of Little Round Lake have not been made. Preliminary studies have shown that the maximum depth is about 16 m, and the mid-summer level of the thermocline about 7-8 m. Pond H is approximately 30 × 33 m in area, with a maximum depth of 2 m. The pond was constructed in 1949 and stocked in 1952 with large-mouth bass, and one year later with blue-gills (*L. macrochirus*).

The lake sediments were collected with a 15 × 15 cm Ekman dredge. Ten samples were taken from Lake Opinicon at 9-10 m depth in the main basin of the lake; 25 samples from Fayetteville Green Lake at 50 m depth; and 23 samples from Little Round Lake at 15-16 m depth. The samples from Little Round Lake were collected in two series, one (samples 1-12) from the central part of the deep-water plateau, and the other (samples 13-23) from the southern edge of the deep-water plateau. It was necessary to use an orange-peel

[1] Dr. Winnifred Frost (personal communication) is currently examining salmonid scales and vertebrae found in a sediment core taken from Esthwaite Water, England.

sampler to penetrate the silty sediments of Pond H. Twenty-five samples were collected.

The lake sediment samples were sieved through an 80 mesh (to the inch) brass screen with openings of 0.2 × 0.4 mm. The sediment from Pond H was sieved through a coarser screen with openings of 1.1 × 1.4 mm. The individual samples from Lake Opinicon and Little Round Lake were examined separately, whereas those from Fayetteville Green Lake were combined before sieving as was the case for Pond H. The materials retained by the sieves were examined under water in a white enamelled tray with a 6X hand lens. For closer examination of some of the fish remains a binocular microscope (magnification up to 120X) was used.

The sediments of Fayetteville Green Lake and Lake Opinicon did not contain any macroscopic remains of fishes and will not be discussed further.

LITTLE ROUND LAKE

The results for Little Round Lake are summarized in table 1. Most of the scales and bones were identified as belonging to the yellow perch, *Perca flavescens*, a known inhabitant of the lake.

Three general types of scale were present: (a) ctenoid scales resembling those of *Perca flavescens*, with only marginal ctenii; (b) ctenoid scales resembling those of centrarchid fishes, with the ctenii extending anteriorly to the focus of the scale; and (c) small cycloid scales. The cycloid scales might have originated from brook trout, *Salvelinus fontinalis*, that have been stocked in the lake. The scales were identified with the aid of Lagler's "Lepidological studies I" (1947) and by comparison with the scales of known fishes. Both the bones and scales were excellently preserved, and of a beautiful amber color. There was no evidence of muscle tissue attached to the bones, indicating death of the fish some (unknown) time previously.

The manner in which the scales and bones are distributed over the floor of the lake is a question of prime importance. In rare samples, such as no. 6 and no. 14, most of the bones (disarticulated) of separate perch heads were found. The skull bones had apparently remained more or less in place. Most of the other samples showed rather peculiar distributions of parts indicating that the various fish remains had been to some extent scattered around the lake bottom. An intact perch has roughly 2700 scales, 40 vertebrae, 40 neural arches, 40 ribs, and 20 hemal arches, i.e. a ratio of 68 scales: 1 vertebra: 1.5 arches (neural and hemal): 1.0 ribs. The corresponding ratio for the combined sediment samples was 24 scales: 1 vertebra: 0.6 arches: 0.3 ribs. The ratios of scales: vertebrae and ribs: vertebrae were found to differ significantly ($P << 0.01$) from the ratios expected of an intact perch, but the scale: rib ratio showed no significant difference. The above data together with the manner of occurrence of the ribs suggest selective sorting and/or decomposition of the fish parts.

Correlation coefficients (r's) were determined for scales and vertebrae and for scales and arches (neural and hemal), separately for the two series of samples as well as for the combined set of 23 samples. The relevant data are presented in table 2. The correlation coefficient for vertebrae and arches was

not determined because in many cases the neural and hemal arches were still attached to the vertebrae. The negative values of $r_{\text{scales-vertebrae}}$ and $r_{\text{scales-arches}}$ for samples 1-12 certainly do not support a hypothesis of a strong marked positive correlation. In samples 13-23 there was a significant ($P = .05$) correlation between scales and vertebrae, but when all samples were combined no statistically significant relationships emerged.

In one sample (no. 10) a complete examination was made of all the scales using a binocular microscope with a magnification up to 120X. Of 148 scales determinable as to type, 134 were ctenoid and 14 cycloid. Most of the ctenoid scales were from *Perca flavescens*; however, it was impossible to make out the distinguishing characteristics on the smaller scales, and some of these may have been derived from other species of fish. Table 3 shows the frequencies of the different size classes of ctenoid scales in the sample. The smallest scales were of about the same dimensions as the openings of the sieve. Of 139 scales that were determinable as to age, 128 were definitely from fish in their first year of life, and 6, 4, and 1, probably from fish in their second, third, and fourth years of life, respectively. The annuli showed no tendency to be near the edges of the scales, suggesting that the mortality was not the result of a *late* winter kill. It would be fallacious to assume that the age distribution of the scales corresponds to that of the fishes that gave rise to the scales. This would be justified only in the absence of sorting and selective decomposion, or, in the presence of sorting if the entire lake had been randomly sampled.

In addition to fish remains the samples contained large numbers of cladoceran ephippia; some unidentified, amber-colored, doughnut-shaped objects that ranged up to 1 mm in diameter and were about one-half as abundant as the scales; occasional twigs and leaves of both deciduous and gymnospermous trees; and a few floatoblasts of the bryozoan *Cristatella*. Living chironomid larvae were not observed in any of the 23 samples. This notable fact suggests a severe oxygen depletion in the bottom waters, and may thus be indirectly correlated with the abundance of fish remains. Little more can be said until the detailed limnology of the lake is known.

POND H[2]

From the records kept by Cornell Fish Hatchery workers it was clear that approximately 350 centrarchid fish (large-mouth bass and blue-gills) had "disappeared" in the pond in the two years prior to the time of sampling (November, 1958). That is to say, this number of fish was not accounted for in terms of stocking, repeated population cenuses, and human removal. It was assumed that these fish died in the pond, and that their scales were not removed from the pond.

The orange-peel sampler penetrated the sediment to an average depth of 2-3 cm. This depth corresponds roughly to the annual thickness of sediment deposited in the pond. We therefore expected to find the remains of about 175 centrarchids in the uppermost 2-3 cm of the entire pond. The number of scales

[2] This section is written in collaboration with Mr. George Tanner.

on a bass or blue-gill is slightly over 2,000. The total area of the pond is about 1,000 m^2. The combined 25 dredge samples covered an area of 1.75 m^2, i.e. about 0.175 percent of the surface area of the pond. *Assuming* that the sampling was random with respect to the scales in the sediment and that no destruction of scales had occurred, we expected to find 610 centrarchid scales in the total of 25 dredge samples. Actually only 26 centrarchid scales were found, suggesting that only about 6 of every 100 original scales had been preserved. Two considerations suggest that this estimate is low: (a) some of the blue-gills were doubtless eaten by the bass, and their scales destroyed in the gastro-intestinal tract of the predator; and (b) the coarse-mesh sieve probably retained only the larger scales present in the sediment. In view of a number of uncertainties that are involved in the calculation, the value of 6 percent is taken only to indicate the general order of magnitude. The preservation in this case pertains to a period of one year in a pond with a relatively high rate of sedimentation. The short time and rapid burial would both tend to favor scale preservation.

In addition to the 26 centrarchid scales isolated from the sediment of this pond, nine other scales were found, one of which appeared to have come from the minnow *Notemigonus crysoleucas*. Thirty-six opercula of the snail *Viviparus contectoides*, a known inhabitant of the pond, were also found. A single pectoral spine of the channel catfish, *Ictalurus punctatus*, was the only non-scaley fish part that was found in the collected sediment. The catfish (or the spine) must have entered from Cascadilla Creek which supplies the water for the pond.

DISCUSSION AND CONCLUSIONS

The above results indicate that fish remains do occur in some lake and pond sediments; however, the data are inadequate to formulate any general statement on the probability of finding such remains in any particular body of water. Attention was focussed on deep water sediments because it is in such locations that limnologists are prone to take sediment cores. Shallow water lacustrine sediments may contain more abundant remains of fishes. It would appear that in Little Round Lake, some specific and perhaps catastrophic phenomenon was responsible for the death of the perch. In the case of Pond H it is possible that some of the scales entered the pond as such from the waters of Cascadilla Creek; however, in such a case one would not have expected such a high proportion of centrarchid scales. It seems more likely that these were derived tfrom the bass and blue-gills resident in the pond.

In closing, it is instructive to calculate the frequency with which various fish parts might be expected to occur in the uppermost 15 cm (the depth to which the dredge penetrated the sediment) of a core from Little Round Lake. Given a corer of 4 cm diameter, one would expect to find an average of 19 scales, 0.5 vertebrae, and 0.5 neural and hemal arches, with occasional evidence of other bones in the top 15 cm. The conclusion from this preliminary work is that fish remains can be expected to be found in cores of lake sediments, but only in certain lakes and perhaps only during certain stages of a lake's history.

TABLE 1

Numbers of scales, vertebrae, neural and hemal arches, and ribs in 23 Ekman dredge (15 × 15 cm) samples from the deep water sediments of Little Round Lake, Ontario (see text for explanation).

Sample	Scales	Vertebrae	Neural and Hemal Arches	Ribs	Other
1	282	4	4	0	
2	271	6	3	0	
3	185	16	11	0	
4	147	13	4	0	
5	108	25	7	0	
6	173	8	5	35	Skull and opercular bones abundant
7	154	10	4	1	
8	273	10	1	1	
9	175	16	1	0	
10	180	72	41	0	
11	153	2	3	0	
12	497	7	8	0	
Av. 1-12	217	16	8	3	
13	228	12	6	2	
14	898	15	4	30	Skull and opercular bones abundant
15	1345	40	37	35	
16	630	8	3	0	
17	291	8	1	0	
18	376	8	3	0	
19	271	8	6	1	Some opercular bones
20	231	10	5	0	
21	267	3	5	0	
22	500	33	43	0	
23	667	5	2	1	
Av. 13-23	519	14	10	6	
Av. 1-23	361	15	9	5	

TABLE 2

Correlation coefficients for scales-vertebrae and scales-arches (neural and hemal) in two series of sediment samples from Little Round Lake, Ontario.

Correlation between:	r (samples 1-12)	significant at $P = .05$?	r (samples 13-23)	significant at $P = .05$?	r (samples 1-23)	significant at $P = .05$?
scales-vertebrae	−.23	no	+.67	yes*	+.19	no
scales-arches (n. & h.)	−.07	no	+.49	no	+.36	no

* Just significant at $P = .05$; not significant at $P = .02$.

TABLE 3

Frequencies of scale size-classes in sample #10 from Little Round Lake, Ontario. The distance from the focus to the anterior margin was determined for each scale, and the scales then grouped according to successively increasing 0.16 mm size intervals. The scales in size class 1 (0.00 – 0.16 mm) would have been too small to be retained by the sieve used to concentrate the scales.

size class	frequency	size class	frequency
1	.00	9	.04
2	.38	10	.00
3	.27	11	.00
4	.11	12	.00
5	.09	13	.00
6	.03	14	.01
7	.04	15	.01
8	.02		1.00

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DETERMINING DIRECTION OF FLOW IN BASALTS

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ABSTRACT. A number of structural features can be used to determine the direction of advance of individual lava flows in the great accumulations of flood basalts such as form the Columbia River plateau and the lava platform of the Oregon Cascades. Three are particularly valuable, however, because they are abundant, easily interpreted, and give unambiguous results: (1) Inclined columnar joints in flow interiors pitch toward the source of the lava; (2) Spiracles (also pipe vesicles and vesicle cylinders) rise from the bottoms of flows and bend or trail out downstream; and (3) Primary foreset bedding develops in most palagonite complexes.

INTRODUCTION

In studies of flood basalts (Tyrrell, 1937), such as those that cover the Columbia River region, determination of the direction of advance of individual lava flows is of importance in gathering data on provenance, position of eruptive centers, petrologic variations, and many other problems. This paper describes methods that proved useful in establishing the direction of flow in the Miocene to Recent basaltic lavas of the Pacific Northwest. Two markedly different kinds of basalt are found in this region: (1) The olivine-poor to olivine-free Columbia River basalt; and (2) The olivine-rich basalt of the Cascade Mountains and plateaus of south-central Oregon.

The Columbia River basalt is chiefly of Miocene age, but locally extends into the early Pliocene. This basalt occurs in thick flows, many over 150 feet and few less than 30 feet in thickness, and it reached the surface in huge fissure eruptions. The fissures through which the lava rose are grouped in a few large dike swarms separated by scores or hundreds of miles of lava field in which no dikes appear. These dike swarms are not associated with volcanoes. In general the Columbia River basalt is a monotonous series of lavas with little petrologic variation, although there are slight differences in texture, mineralogy, and chemical composition that characterize the lavas from different dike swarms.

The olivine-rich basalt of the Cascade Mountains and plateaus of south-central Oregon is chiefly Pliocene to Recent, and in many places rests with marked unconformity on the Columbia River basalt. It occurs mainly in thin lava sheets averaging less than 20 feet thick. In contrast to the Columbia River lavas it was erupted chiefly from low shield volcanoes. The lavas emerged in part as summit extrusions, but mainly in fissure eruptions that broke forth on the flanks and at the base of the lava cones. In places huge fissure eruptions of the Columbia River type have also contributed to building up the olivine basalt lava fields, but in contrast to the Columbia River dikes most of the olivine basalt feeders are surmounted by a line of cinder cones and low lava shields built during the waning stages of vulcanicity. As might be expected from their local concentration into many small and perhaps shallow magmatic hearths, the olivine basalts show considerably more variation in texture and composition than does the Columbia River basalt.

More information on the characteristics of these two series of basalts is given by Waters (1955). For the present discussion an important point is that the Columbia River basalt was erupted in huge flows of relatively "dry" and

perhaps slightly more viscous magma that traveled to great distances (for some flows at least 50 miles from the feeding source), whereas the olivine basalt magma was richer in volatiles, and it flowed chiefly in thin lava streams and overlapping tongues. Even the large fissure eruptions of olivine basalt did not retain individuality as single flows, but subdivided into numerous flow units (Nichols, 1936) and broke up into complicated sets of internal lava tubes similar to those described by Skeats and James (1937).

This paper is not concerned with flow directions on recent shield volcanoes where the relations of individual flows to vents are generally obvious from their morphological form and minor surface features. The attempt is to find reliable criteria for determining direction of flow in older dissected lava fields, and especially in the huge accumulations of flood basalts. Nevertheless, in the search for criteria, many flows were studied whose source and direction of spread is known. Some of these are Pleistocene olivine basalts that flowed down former stream valleys and have since been trenched and well exposed by reorganization of the streams they displaced. Hundreds of such partly eroded intercanyon flows fill valleys on the east slope of the Cascade Mountains in Oregon and southern Washington. Broad sheets of Columbia River basalt which spread over flat ground from known vents were also used, as well as a few intercanyon flows in this formation. Many structures were found that indicate direction of lava advance, but only a few are sufficiently common and diagnostic to be of general use. Among the best are inclined columns in the interior of flows, spiracles that rise from the bases of flows, and certain features of pillow lava-palagonite complexes. The use of these features to determine direction of flow seems to be little known, so they are described fully. Other criteria, better known, are indicated by citation to pertinent references, and less valuable criteria are mentioned only briefly.

STRUCTURES OF FLOW INTERIORS

Inclined columns in flow centers.—The principal factors that control the size and shape of columnar joints in basalt have been well stated by James (1920). Columns grow at right angles to isothermal surfaces within the cooling lava, but if the lava flow is thin, or if cooling is rapid and irregular, columns may not form. Instead the lava breaks into hackly fragments or joints irregularly. Many joints are vertical, however, and so simulate columnar structure if observed only in cross section.

Most of the thick flows of Columbia River basalt show excellent two-tier columnar jointing: a lower tier of coarse but generally well shaped columns grows upward from the base, while a longer tier of thin and more irregular columns sprouts downward from beneath the flow top. The two meet along a prominent horizontal parting about two-thirds of the distance from the top of the flow. Under normal conditions of static crystallization (fig. 1, left) isothermal surfaces are nearly horizontal, and so the two tiers stand vertical. Thus in figure 1, left, lava at points A and B has reached the same temperature, columnar joints are growing in it, and they continue to grow vertically because the isothermal surfaces between A and B are essentially horizontal.

But if forward movement of the lava is resumed after columnar joints

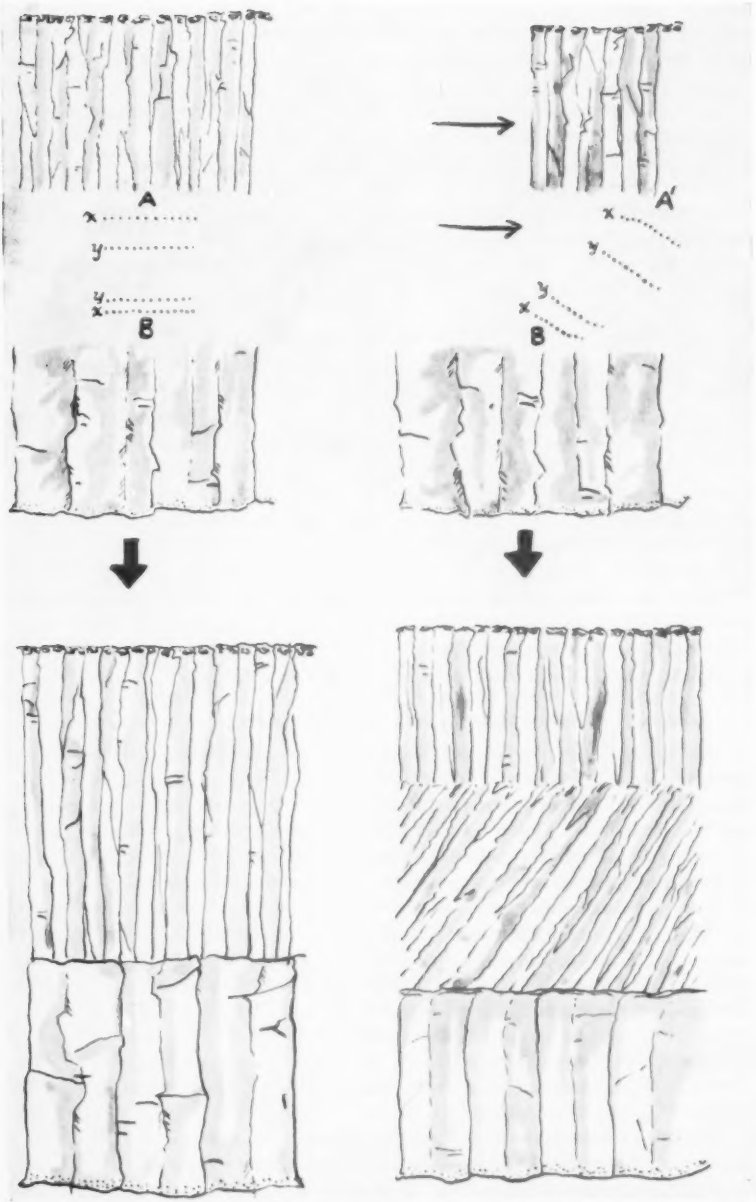

Fig. 1. Cross sections of lava flows showing the growth of two-tier columns under static conditions (left) and with renewed movement after partial solidification (right). Dotted lines x, y, in upper sketches represent isothermal surfaces.

have grown some distance from the top and bottom of the flow the isothermal surfaces will be tilted because of transport of hotter lava from nearer the vent over the already solidified lower tier of columns. For example, in figure 1, right, creep of the still liquid interior has carried A to A′. Now two similarly related points of the same temperature (A′ and B) are no longer in the same vertical line, and in consequence the isothermal surfaces between them are inclined. Continued growth of both the basal and the top tier of columns, of course, occurs at right angles to these tilted isothermal surfaces. Hence new extensions on both the upper and lower tier of columns will be tilted (fig. 1, lower right, and plate 1, fig. 1). The columns appear to have been dragged in the direction of flow, but of course the "drag" is not caused by actual bending or distortion of already solidified columns: it is only the result of continued growth of the joints in conformity with the tilted isothermal surfaces.

In the thick flows of Columbia River basalt this structure is common, and it affords one of the most reliable indicators of direction of flow. Moreover, because it occurs in flow centers, it is generally well exposed on the cliff faces so common in eroded basaltic lavas.

In places confusion may arise because inclined columns can be formed in other ways. They sprout from the irregular topographic surface if a flow buries hills or cliffs. If water from an adjacent stream pours over the surface of a lava flow before the lava completes its solidification, a jumble of fan-like columns may radiate in all directions from numerous chilled centers at the top of the flow. But such inclined columns grow from the base or top of flows, whereas the tilted columns useful in determining direction of flow are confined to flow centers.

In contrast to the Columbia River basalt, columnar jointing is not common in the thin flows of olivine basalt, but it generally appears in flows over 25 feet thick and is almost ubiquitous in thick flows that have ponded in topographic depressions. Most large olivine basalt flows, however, have subdivided into flow units characterized by irregular vertical joints and slab-like horizontal partings. In places columnar joints have sprouted from the contact between flow units, and under appropriate conditions of cooling and movement the joints curve forward in the direction of lava advance. But such columns are rare, and they do not have the uniformity in trend that characterizes columns from the interior of thick two-tier flows.

Filled lava tubes.—Many of the olivine basalts contain numerous filled lava tubes—these are rare in the Columbia River basalt. The trend of the lava tubes, and their internal jointing, afford reliable evidence of the direction of lava advance.

Some lava tubes are underground continuations of lava rivers that emerge from vents, pour down steep slopes, and gradually crust over and disappear into the flows they fed. Other tubes form by the protrusion of finger-like extensions and broader tongue-like flow units from the front of advancing flows (Skeats and James, 1937, p. 253, 256). Although most accounts of historic eruptions describe only the relatively small amoeboid protrusions that squirt from the toes of advancing flows, Wentworth and Macdonald (1953, p. 33) report examples "several tens of feet in diameter", and also record that "In

cross section a flow unit often consists of a large number of small, filled tubes." In the olivine basalts of the Pacific Northwest, as in those of western Victoria (Skeats and James, 1937), large lava tubes are common within flow units. They occur singly, clustered into groups, or may grade into large flat tongues and flow units.

Most of the literature on lava tubes deals with open lava caves and tunnels. These are common on steep slopes where the lava retains enough head to breach the front of the flow and drain after eruption ceases. But in extensive lava fields that have spread over more gentle slopes the filling of the tube generally remains in place and hardens within its host. Moreover, in a complex field of flood basalts the lava tubes, even if drained, are likely to be inundated by later flows as in the interesting examples described by Wentworth and Macdonald (1953, p. 27) which occurred during the 1935 and 1942 eruptions of Mauna Loa. In Oregon and Washington filled lava tubes are much more common than open ones, but because they are far less conspicuous in outcrop they are easily overlooked.

One of the most common kinds of tube filling is the "War Bonnet"—a local name given to a conspicuous outcrop formed by the cross section of a tube filling that appears high on a cliff facing Columbia River at White Salmon, Washington. Figure 2 is a sketch of two large "war bonnets" from a

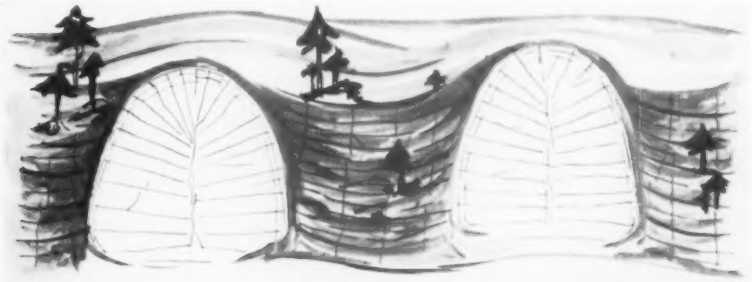

Fig. 2. Lava tube fillings in an olivine basalt near Troutdale, Oregon. Note the columnar joints radiating from the top and walls of the tubes, and the slab-like roughly horizontal parting of the flow in which they are embedded. Each tube is about 35 feet high.

cluster of 5 tube fillings in an olivine basalt flow near Troutdale, Oregon. Columnar joints grow inward from the roof and walls of the tubes, giving a radial structure that resembles the feathers on an Indian headress (pl. 2, fig. 3). Columnar joints are typical of tube fillings even though the rock surrounding them rarely shows good columns. Insulated by its shell of non-conducting older basalt, the filling of a lava tube cools slowly and uniformly as in a dike or small sill. Indeed the presence of good columnar jointing in small masses of olivine basalt is an indication that the mass is either a tube filling or an intrusive. Renewal of movement during cooling of a lava tube causes "bent" columns of the kind found in the centers of two-tier flows.

Another kind of tube filling is not columnar. It is characterized by curving platy joints or by bands of vesicles arranged in concentric zones

parallel with the walls of the tube. Although comparatively rare in the basalts of Oregon and Washington such tube fillings are evidently more common in Hawaii. Excellent pictures of them have been published by Wentworth and Macdonald (1953, fig. 23, p. 46) and by Palmer (1929, figs. 1 and 2, p. 272, 273).

Structures preserved near fronts of laterally spreading flows.—Near the fronts of a few thick flows of Columbia River basalt a series of prominent vertical fractures were observed which trend roughly parallel with the direction of lava motion (plate 1, fig. 2). These are tension cracks, similar in orientation to the longitudinal crevasses that develop in the front of a glacier. In an intercanyon flow of olivine basalt along White Salmon river near Husum, Washington, transverse "crevasses" were noted that spring from the walls of the flow and curve upstream toward the source of the lava like those in a valley glacier. Such tension cracks in lava are easily distinguished from post-solidification tectonic faults and fractures because they are restricted to the upper part of the lava flow—they do not cut the lower tier of columnar joints—and they also act as cooling surfaces and deflect the columnar joints toward them. Both these relations are shown in the flow illustrated in plate 1, figure 2.

Although these fractures are useful in determining the direction of the source of lava flow, they are rare. Brinkmann (1957), however, has made the interesting observation that the same stress pattern which is responsible for the development of these tension fractures also causes a distinct and measurable distortion in the hexagonal cross sections of the normal upright columnar joints. The hexagons appear somewhat flattened, and their longer axes are aligned parallel to the trend of potential crevasses like those described above. Brinkmann (1957, p. 532) explains:

> Wenden wir diese Betrachtungen auf die natürlichen Verhältnisse, die Schrumpfungsspaltenmuster vulkanischer oder subvulkanischer Gebilde an, so gelangt man zu einleuchtenden Ergebnissen. Die Übereinstimmung des Spaltenbildes eines Talgletschers mit der Anordnung der langen Achsen der Summenpolygone auf einem Lavastrom besagt, dass grundsätzlich auch in der fliessenden Lava Rand- und Querspalten entstehen. Sie sind aber von dem hexagonalen Netz der Schrumpfklüfte überlagert und kommen nur in einer Verzerrung der Sechsseite zum Ausdruck.

Thus it appears that Brinkmann's method of plotting the ellipticity of a columnar joint pattern will yield much more abundant data than can be obtained from the much rarer longitudinal crevasses.

STRUCTURES OF FLOW TOPS

Most of the flows of Columbia River basalt show typical aa tops. (For descriptions, and a discussion of the origin of various kinds of flow tops see Dutton, 1884, and Macdonald, 1933). The surface of the flow is covered, to a depth of from a few inches to over 15 feet, with jagged fragments of granulated lava and bits of glassy clinker injected from below by spiny broken slivers and steeply imbricated slabs of the underlying massive lava. Such surfaces, even where well exposed, do not give reliable evidence of the direction of lava motion. In places a general idea of the direction of flow may be gained by study of the orientation of the steep slabs that inject the cindery

tops from below. The long dimensions of the slabs are commonly (but not uniformly) about at right angles to the direction of flow, and the steeply imbricated slabs generally (but not always) dip toward the source of the lava. However, the relations of these splintery slabs are in many places so complex and irregular, even within a small area, that little reliance can be placed on such measurements. A few scattered observations are useless.

A few of the Columbia River flows, and perhaps 30 percent of the olivine basalts, have tops of pahoehoe. Pahoehoe surfaces show abundant minor features—ropes and cords, lobate ridges, stretched vesicles, pulled glass filaments, broken and distorted bubbles, skid marks, lava flutings, pressure ridges and pressure cracks, schollendomes, "barriers," and many others—that have formed by the stretching and piling up of viscous lava as it crusts over and is swelled upward by pressure from below or is dragged forward by the moving lava beneath. Detailed descriptions of surface features on historic and Recent flows have been given by many observers. A particularly concise and well-illustrated analysis with appropriate references to older literature is by Wentworth and MacDonald (1953, p. 32-65).

Although nearly all such surface features are the product of lava motion, their interpretation is not without ambiguity. Ropy folds of lava may extend either parallel or at right angles to the direction of motion (Russell, 1902, p. 92-93; Wentworth and Macdonald, 1953, p. 35), or they may drape forward in lobate festoons. Furthermore they are twisted and cross folded, or broken into blocks by continued movement. In many flows broken slabby crusts of pahoehoe a few inches thick have been piled up in complicated block fields of steeply-tilted slabs (Macdonald, 1953, p. 175; Krauskopf, 1948, p. 1281-1282). Even coarser features such as pressure ridges and "barriers" may lie either parallel or transverse to the direction of flow, and their mode of origin is vigorously debated (cf. Russell, 1902, p. 94-97; Skeats and James, 1937, p. 255-260, 270-272; and Nichols, 1939).

Surface features soon disappear by weathering, and in superposed flows they are generally concealed under blocks that have slid down from the flow above. For these reasons oriented surface features have been little used in this study, but might prove of value if well exposed so that many observations could be made and the results analysed by the statistical methods currently applied to directional structures in sedimentary rocks (Pellitier, 1958; Potter and Olson, 1954). From limited observation the writer is of the opinion that measurements of the trend of stretched bubbles and pulled glass filaments on the common filamented pahoehoe (Wentworth and Macdonald, 1953, p. 35, 37), and the plotting of the poles of oriented slabs of massive lava that form injections beneath aa tops will give the least ambiguous results of the many kinds of structures that occur on flow tops.

STRUCTURES OF FLOW BOTTOMS

Spiracles and pipe vesicles.—Water vapor and other gases surge upward into liquid lava that covers marshy ground. Trapped pockets of air, gas from burning vegetation, and steam expelled from soils and water-soaked sediment may also boil upward into the flow. These vapors chill the lava, freezing it to

basalt glass along the margins of the jets of escaping gas. If the chilling is rapid the gas channels may be preserved intact when the flow congeals (figs. 3 and 4; and plate 2, fig. 2).

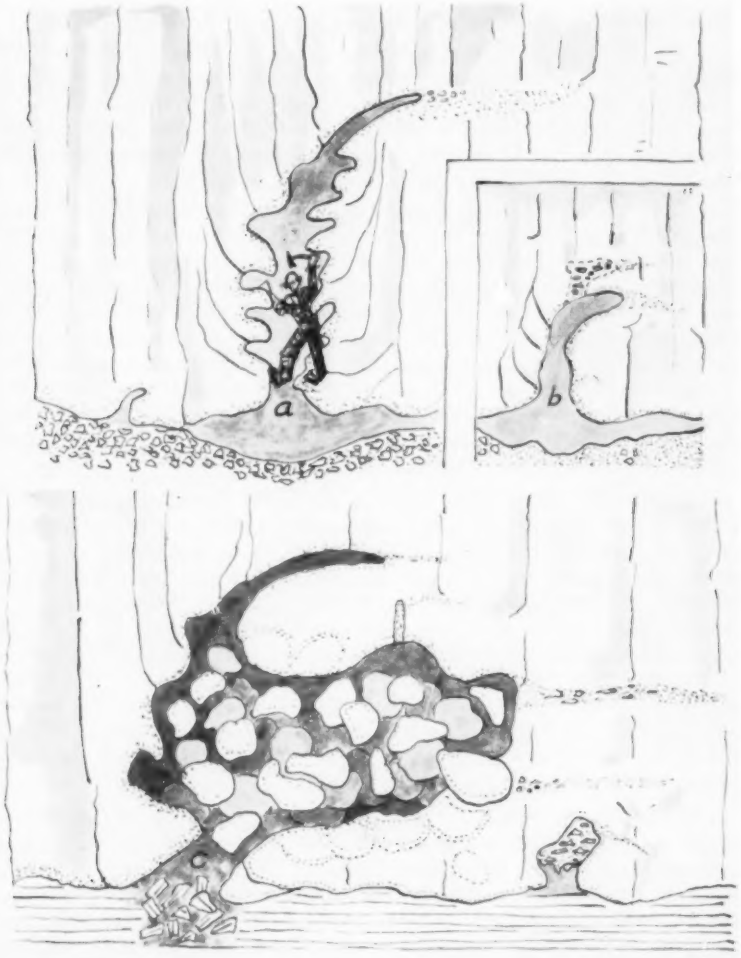

Fig. 3. Large spiracles in the Columbia River basalt. In each example the direction of flow is to the right. a). Large spiracle with lobate walls rising from a thick mass of aa clinker. Moses Coulee, Washington. b). Spiracle, with tongue-like clusters of vesicles, rising from a pit in the underlying pahoehoe flow top. Near Olex, Oregon. c). Large spiracle rising from beds of water-laid tuff, and expanding into a cavity filled with pillows. Note smaller spiracles, vesicle clusters, and vesicle cylinders extending from the edges of the cavity. Clackamas River, Oregon.

The upper ends of most gas chimneys and tubules bend in the direction of lava motion, but some break up into irregular cavities and clusters of vesicles that trail out horizontally downstream. Steam explosions in water saturated sediments beneath the flow blast mud and other debris into the chimneys; this, too, is scattered to leeward by lava movement.

The large gas chimneys are called *spiracles* (Fuller, 1931, p. 292-293, 299). In the Columbia River basalt—where they are common—spiracles range from a few inches to scores of feet in diameter. Most widen slightly or even mushroom after rising a few feet into the flow. After rising vertically for as little as a foot or as much as forty feet most of them bend abruptly or else trail out as vesicle clusters in the direction of flow.

The walls of the larger spiracles are intensely corrugated by horizontal lobes and deep flutings—probably formed as the lava attempts to pour into the spiracle but is continually pushed back by the rising column of steam. Upon these corrugations droplets and larger masses of liquid lava have spilled and congealed. Because of the deep horizontal corrugations it is possible to climb up the inside of some of the larger spiracles, using the corrugations on the sides like the rungs on a ladder (fig. 3a). Spiracles are nearly circular in cross section, but as they bend forward in the direction of flow they flatten, lose most of their corrugations—or even acquire a linear fluting—and break up into strings of vesicles. Some subdivide upward into several tongue-like patches of vesicles separated by massive lava (fig. 3b).

Most small spiracles are open, or are partly filled with fragments of sediment shot up from below and with pieces of glass that have dripped or exploded from the walls above. Nearly all spiracles more than 10 feet in diameter are partly filled with pillow lava and palagonitized glass breccia (fig. 3c). Some pillows are attached to the lobes on the walls but most are free. The core of loose pillows may rise vertically for several feet and then be rolled forward in the direction of lava advance, or more commonly the pillows are drawn forward directly from the base of the flow, ascending at an angle of about 40 degrees over a ramp of complexly intermingled pillows and massive lava. From the top of these pillowed masses small spiracles and vesicle cylinders (Du Toit, 1907, p. 15) burst upward into the overlying lava, and clusters of vesicles string out from their downstream sides.

Careful search of the base of nearly any large flow of Columbia River basalt—especially in areas where it covers an old soil, ash zone, or layer of sediment—will nearly always reveal spiracles. Therefore they are one of the most useful features from which to determine direction of flow for this formation. Spiracles are also common in the olivine basalts, but because these flows are rarely more than 20 feet thick the large complex spiracles that contain pillows are rare (although areas of pillow lava are common). Spiracles in olivine basalts are generally less than one foot in diameter, and seldom more than 3 feet high. They grade downward in size into the well-known *pipe vesicles* whose use in determining direction of flow has long been known (Du Toit, 1907; Walker and Parsons, 1922; Butler and Burbank, 1929, p. 27; Fuller, 1939, p. 305).

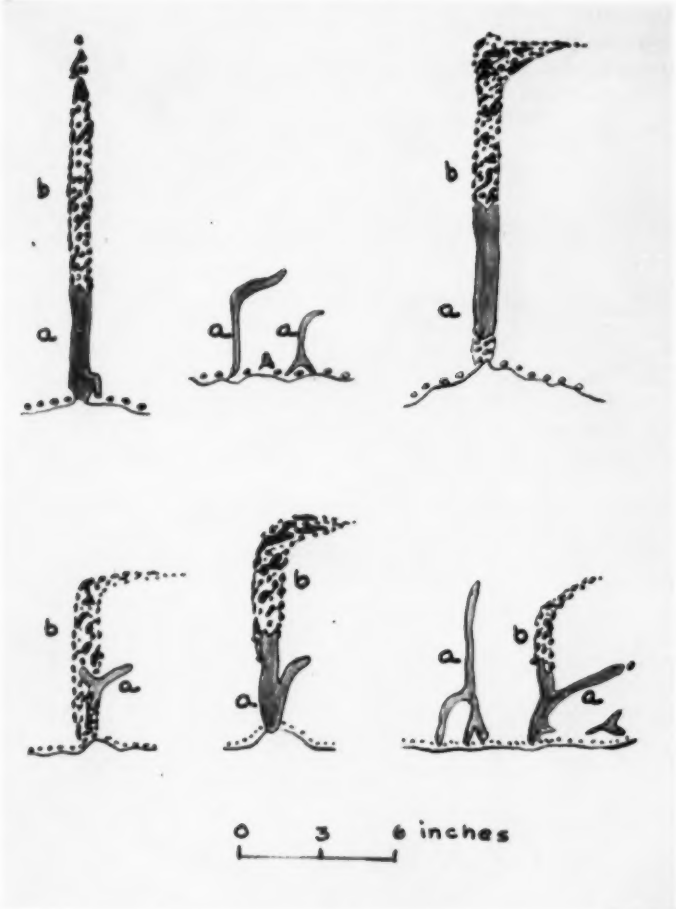

Fig. 4. Pipe vesicles (a) and vesicle cylinders (b) rising from the base of olivine basalt flows near Hood River, Oregon.

In the olivine basalts of Oregon and Washington pipe vesicles tend to occur in clusters, and, as in the Drakensberg (Du Toit, 1907, p. 15), they are generally associated with vesicle cylinders (fig. 4). Unlike spiracles, pipe vesicles and vesicle cylinders are not notably more abundant in flows that cover soil or sediment than they are in other flows. Pipe vesicles are not common in the Columbia River basalt: aparently the bases of the olivine basalts were highly fluid and easily penetrated by even small pockets of air trapped in irregularities beneath the flows, whereas the Columbia River flows appear to have had a tough hide which in most places could be ruptured only

by steam released with explosive violence from pockets of water or patches of water-soaked sediments.

Smeared vesicles.—In most descriptions of moving aa lava it is stated that as the flow creeps forward broken rubble from the top of the lava cascades down the end of the flow and is rolled under the advancing margin in "caterpillar tread" fashion. Theoretically, therefore, the base of a flow should consist of a layer of granulated clinker and slag like that on the top of the flow. This expectation is not confirmed by observations of either the flows of Columbia River basalt or of the olivine basalt of the Cascades. Most flow bottoms show a complete absence of breccia and rubble, others reveal only scattered discontinuous patches. Even flows that have over 15 feet of breccia at their tops show little or none along their bases. Butler and Burbank (1929, p. 27) also noted that breccia is typically absent at the base of the Keweenawan flows in Michigan although thick layers of rubble formed on their tops. Apparently in these voluminous flows the front advanced rapidly and was sufficiently "live" during most of the period of lava movement to allow the liquid lava to break forth from beneath the surface rubble and outdistance its blocky cover.

The base of a typical flow of Columbia River basalt is flat or irregularly lobate, conforming roughly to the shape of the ground on which it rests. A chill zone of massive lava gives way in an inch or two to a slightly vesicular layer containing numerous small round bubbles. Upward the vesicles become slightly larger but less abundant. In most flows this vesicular bottom grades upward into massive non-vesicular lava within less than three feet. The vesicles within two or three inches of the base are generally spherical, but higher in the flow they may be flattened into pancakes parallel with the base of the flow.

Where the base of the flow meets an obstruction in its course—either rubble rolled under the front, or more commonly an irregularity in the land surface covered by the flow, the flattened vesicles behind the obstruction are tilted upward so that they dip toward the source of the lava. Along the top of the upstream end of the obstruction they are commonly smeared into long flattened spindles strung out in the direction of flow. These flattened and spindle-shaped vesicles are generally reliable indicators of the direction of lava advance.

Ambiguities can arise in the interpretation of smeared vesicles. If the top of the obstruction is irregularly jagged the flow lines will diverge around the peaks, causing difficulty in ascertaining the regional direction. Similarly the tendency of the basal part of the lava to spread sideways on first meeting an obstruction can cause confusion unless only those vesicles near the top of the

PLATE 1

Inclined columns and longitudinal crevasses in Columbia River basalts

Fig. 1. Inclined columns in the center of a flow of Columbia River basalt. The tilted columns merge both upward and downward into the normal vertical columns. The direction of flow was to the right, but with a component toward the camera. Near Cottonwood, Idaho.

Fig. 2. Longitudinal "crevasses" in the upper part of a 70-foot thick flow of Columbia River basalt. Note that the crevasses do not cut the lower tier of columns, and that they have acted as cooling surfaces during the growth of the upper tier. Crooked River, about 18 miles above Prineville, Oregon.

PLATE 1

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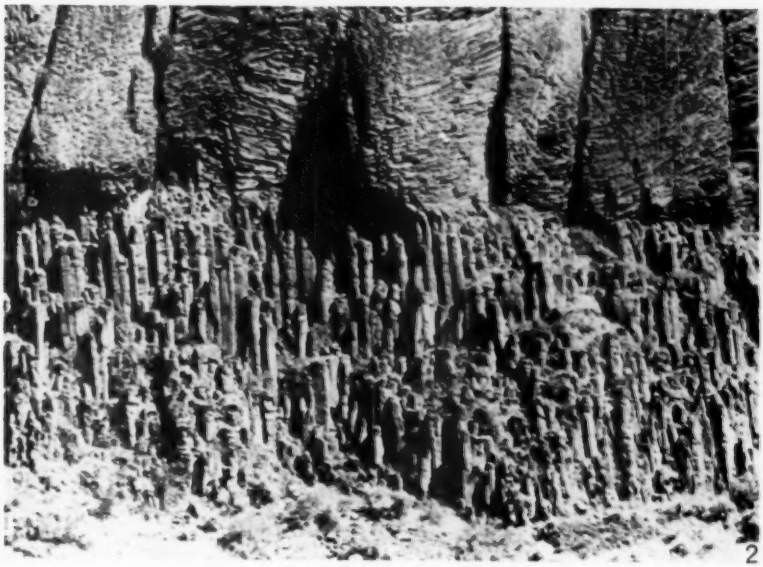

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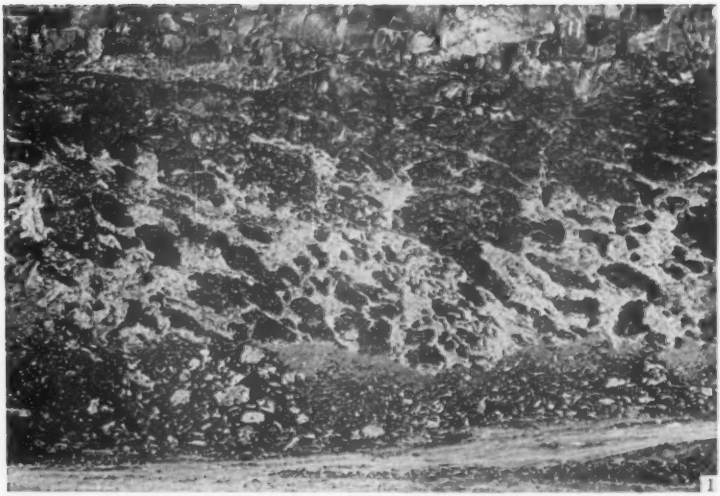
Inclined columns and longitudinal "crevasses" in Columbia River basalts

PLATE 2



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Some structural features of Columbia River basalts

obstruction are used. Furthermore if the obstruction consists of rubble rolled under the front of the flow there is the danger of mistaking a slab of pahoehoe containing stretched vesicles formed on the flow top for the true lava skin at the base of the flow.

Other features of flow bottoms.—Comparatively rare features that may assist in determining the direction of flow are skid marks on either the bottom of the flow or on the ground it covers, distorted and overfolded clays and soils that have been overridden by nearly solid lava, and tree molds in the base of lava flows (clusters of vesicles and streaks of whitish alteration caused by gases released from the wood are strung out from the tree in the direction of flow).

STRUCTURES OF PILLOW LAVA-PALAGONITE COMPLEXES

When flood basalts build an extensive lava plain there is sure to be much disruption of the former drainage. Streams are ponded against the margin of the lava flood and shallow lakes spread widely over the edges of the newly congealed lava. Flows from new eruptions pour into these marginal lakes, filling them with a complex of pillow lavas and granulated basalt glass. The process is repeated again and again. The glass of these subaqueous breccias is nearly always partly altered to yellow earthy palagonite (Peacock and Fuller, 1928). Palagonite, too, is unstable and is generally at least partly decomposed to clay minerals, zeolites, and other products. In older rocks such as the submarie flows of Eocene age in Western Oregon and Washington (Waters, 1955, p. 704-707), the palagonite has generally been completely altered to a sooty black or dark green argillite composed of varying amounts of montmorillonite, nontronite, saponite, celadonite, and other clay minerals, along with chlorite, zeolites, carbonates, opal, and chalcedony.

Pillow lava-palagonite complexes are widespread along the margins of the Columbia River basalt. They show much variation in structure, depending chiefly on the viscosity of the lava and the depth of water. These variations, and also the use of certain structures in determining the direction of flow, have been ably summarized by Fuller (1931). The following section is largely abstracted from his paper, but is supplemented by additional observations that bear on the problem of determining direction of flow.

Foreset bedded breccias.—Fuller (1931, p. 282-284, and fig. 1, p. 283) describes and illustrates an exposure on the north wall of Columbia River three miles below the mouth of Moses Coulee:

> The breccia is formed locally of granular fragments of vesicular and ropy

PLATE 2

Some structural features of Columbia River basalts

Fig. 1. The entry of a flow of Columbia River basalt into a shallow lake has produced foreset bedded palagonite breccia (light color in lower half of photo) which encloses numerous isolated pillows and inclined streaks of lava. This grades upward into massive pillow lava, and (at the top of the picture) into normal columnar jointed lava formed above water level. South abutment of bridge over Columbia River at The Dalles, Oregon.

Fig. 2. A spiracle, about 4 inches in diameter, rising into the base of a basalt flow that rests on diatomaceous silts. Near Kittitas, Washington.

Fig. 3. Lava tube filling (upper center), Deschutes River, south of Maupin, Oregon.

> sideromelane [clear basalt glass], but as a rule it has been almost completely altered to the yellowish palagonite. The most striking characteristic of these fine breccias is their content of coarse masses of lava lying predominantly in diagonal streaks, which are roughly parallel and have a primary dip towards the north of about 30 degrees. The streaks are composed either of solid basalt or of a series of ellipsoidal or slag-like masses, which are selvaged with glass to a depth of at least a half an inch. With palagonitization their margins blend with the surrounding breccia. These dark masses of lava, contrasting strongly with the yellowish matrix, give the horizon a marked resemblance to foreset bedding . . .
>
> A fluid lava on entering a local body of water would tend to granulate like molten slag and would thus form a fine breccia which would accumulate to a depth approximately equal to that of the water. The fine breccia would settle until its surface attained an angle of repose which, owing to the roughness of the fragments, would be relatively steep. If the molten cascade continued to pour into the water, the accumulation of granulated glass would gradually advance like the foreset bedding of a delta. The inclined bedding would be preserved by the thin sheets and the ropy or ellipsoidal masses that failed to granulate. Except for the possible effects of rising steam, the flow would gradually advance on top of these foreset beds as if on dry land.

Such foreset bedded breccias and associated pillow lava complexes (plate 2, fig. 1) are found at hundreds of localities along the margins of the Columbia River basalt, and are also a common feature of the olivine basalts—especially at the base of intercanyon flows that have displaced rivers from their beds. Therefore they are particularly useful in determining the direction of flow.

A few ambiguities may arise in interpretation:

(1). Between Hood River and Cascade Locks several flows of olivine basalt entered the ancestral Columbia River and built complexes of pillow lava and palagonite into it. But this large and swift river was locally able to rework the granulated glass nearly as fast as it poured in: therefore palagonite breccias with primary foreset bedding are mingled with other cross bedded palagonites showing current bedding. In large outcrops, however, it is easy to distinguish the two kinds of cross stratification because the primary foreset beds invariably have thin sheets of lava and clusters of pillows interbedded with the foresets.

(2). In places weighting of the newly formed foresets by advancing massive lava accumulating upon them has caused underwater slides in the breccia with resulting confusing relations between foreset beds, pillow lavas, and massive lavas.

(3). A somewhat similar confusing relationship has been produced at numerous localities among the Eocene pillow lavas of western Oregon and Washington, and indeed seems to be typical of the margins of large submarine eruptions that were extruded into deep water. Apparently many submarine eruptions first build up great piles of palagonite breccia which are then pushed outward, tunneled into, and partially overridden by pillow lavas and by great pods and central pools of massive lava. This massive lava formed, more or less protected from the water by a carapace of palagonite breccia and pillows, during the later stages of the eruption. The pods and pools of massive lava, and even some of the larger pillows, send intrusive tongues into the overlying breccia; they also disrupt its bedding to a chaos of jumbled blocks. Fuller (1931, p. 294-298) describes outcrops at Douglas Creek, Washington, where a flow of Columbia River basalt entered a deep lake, advanced subaqueously beneath its own granulated breccias, and even burrowed into soft sediments

that formed a steep delta front at one side of the lake. Skeats and James (1937, p. 263) describe tongues of lava that entered a lake in western Victoria and "appear to have flowed against the soft beds and pushed them ahead, or even ploughed through and under them." Such relations have been cited (e. g. Lawson, 1914, p. 7) as evidence that pillow lavas are really intrusives—emplaced far underground—but when a heavy mass of lava flows along the bottom of a water body it should be expected to spread, at least locally, into its own pile of granulated breccia and adjacent unconsolidated sediments. The drastic chilling that the lava has undergone testifies to subaqueous, not subterranean consolidation. But in such tangled lava-palagonite-sediment complexes foreset bedding must be used with caution as an indicator of direction of flow.

Directional structures within pillow lavas.—Fuller emphasized that the proportions of foreset bedded breccia and of lava in thin sheets and pillows varies greatly in different foreset bedded complexes. He found many places where such complexes grade laterally into massive sheets of pillow lava containing little or no palagonite breccia. Many of these massive pillowed zones show no traces of foreset inclination, and when they do show foresets the slope is at a much lower angle than in the foreset bedded breccias. Apparently nearly all pillows come to rest on the bottom before they completely solidify. The adjustment to the shape of the floor on which they rest (which produces the typical bun-shaped and balloon-like forms) is done under static conditions.

In some outcrops of pillow lava the long axes of the buns are aligned, and this has been cited as an indication of the direction of flow. For example, Bartrum and Turner (1928, p. 132) state that pillows in the lava at Twilight Bay, New Zealand "are decidedly overturned and elongated in the direction of flow." The present writer's experience, however, is that elongated pillows may be produced in different ways, and their elongation often has no clear relation to direction of flow. Indeed it may change erratically over short distances even within a single mass of pillow lava. Pillows interlayered with foreset bedded palagonites are commonly (but not invariably) elongated with the dip of the foresets; pillows at the top or bottom of the foreset structure, however, may have an entirely different orientation.

The only criterion that seems generally useful is the foreset bedding of the pillowed complex, and in areas that do not contain palagonite breccias this is difficult to detect because of the low slope of the pillow foresets. Moreover the slope of the individual pillows must be compared with the slope of the bottom or top of the sheet of pillow lava, and these are nearly always hummocky and irregular, not planes. One minor structure, however, is helpful in establishing this relation: in the upper part of many pillows is a small cavity with an arched roof and a flat floor (see Fuller, 1931, fig. 3, p. 287), or the pillow may show a series of flat, pancake-like openings stacked one on top of another (fig. 5). These formed while the interiors of the pillows were still liquid—their floors are flat because the lava leveled off under the action of gravity. Therefore if the pillow has congealed in a horizontal position the flat floors of the cavities are parallel with the long axis of the pillow, but if the pillow congealed as part

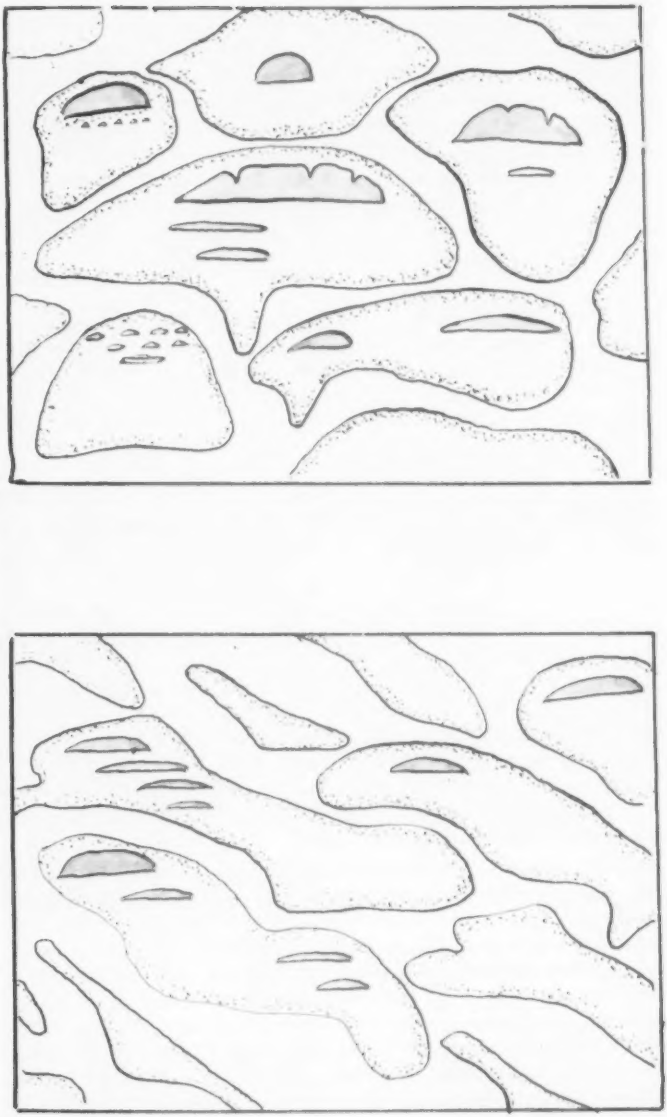

Fig. 5. Cross sections of pillow lavas showing the relation of flat-floored cavities within pillows to pillow elongation in a normal pillowed flow (top) and in a foreset bedded flow (bottom).

of a foreset bedded group of pillows a plane parallel to the base of the pillow will intersect the plane forming the floor of the cavity at an angle (fig. 5).

Spiracles and vesicle cylinders associated with pillow lavas.—Instead of attempting to interpret these elusive structures of pillow lavas one can generally ascertain the direction of flow in a pillow lava complex quickly and certainly from the abundant bent spiracles and vesicle cylinders found in the sheets and pods of massive lava which advanced over a filling of pillows and palagonite, or which spread in pools along the bottom of a deep body of water. Great quantities of steam must rise from a pillow-palagonite complex in process of quenching, as well as from beneath a submarine flow. Most such flows show thick clusters of spiracles, pipe vesicles, and vesicle cylinders.

FLOW STRUCTURES AND LINEATION

The suggestion is often made that a detailed study of the mineral orientation in basalts should give knowledge of the direction of flow, and results of such studies have been published (Brinkmann, 1957, for example). The centers of most basalt flows, however, are massive and show little or no flow structure. Near the base of some flows of Columbia River basalt microphenocrysts of plagioclase form a well-developed platy fabric with (010) predominantly parallel to the base of the flow. Pyroxene, however, generally is not oriented by this flow banding; indeed it may enclose the oriented plagioclase laths ophitically. Flow structures such as these, as well as the better developed ones on flow tops discussed previously, seem of little value in determining direction of flow because of their elusive distribution and because of ambiguities in their interpretation.

CONCLUSION

Inclined columnar joints in flow interiors, spiracles that rise from the bottoms of flows, and primary foreset bedding in palagonite breccias are the most reliable features from which to determine the direction of flow of ancient basaltic lavas.

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THE KEWEENAWAN LAVAS OF LAKE SUPERIOR, AN EXAMPLE OF FLOOD BASALTS*

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ABSTRACT. Individual lava flows of Keweenawan (late Precambrian) age in the Lake Superior basin are characterized by volumes that may exceed 100 cubic miles and by original surface slopes of the order of 10-20 feet per mile or less. In these respects they differ greatly from the flows of shield volcanoes from which so much of our knowledge of lava flowage stems. The volumes are estimated from field measurements. The slopes are estimated by subtracting reasonable initial dips of interbedded conglomerate beds (obtained by comparison with modern fan deposits of similar coarseness) from the angle between top and bottom of flows overlying conglomerate beds, since the lava flows and the gravel-depositing streams are believed to have flowed in opposite directions. Difference in volume alone seems adequate to explain a number of the physical differences between the Keweenawan flood basalts and those of shield volcanoes without appeal to differences in composition, volatile content, or temperature.

INTRODUCTION

As has often been observed, flood-basalt flows differ from the flows of shield volcanoes both in volume of lava and in slope. Since so much of what we know about the flow of lava comes from the keen observations of students of active shield volcanoes, one may be tempted, for lack of comparable observations on active basalt floods, to assume more similarity in mechanism than actually exists. It may be worthwhile, therefore, to make some quantitative comparisons between a group of flood basalts and the flows of shield volcanoes, and to note, if only briefly, some implications of the differences.

The mafic lavas of middle Keweenawan age in the Lake Superior region are one of the world's major accumulations of plateau or flood basalt flows. Estimates of the thickness of the lava series range from 20,000 to 30,000 feet, and the total volume of lava must exceed 100,000 cubic miles. These mafic flows have been arched downward to form a gigantic syncline, the Lake Superior basin or syncline, whose trough is now occupied largely by the lake itself. The rocks generally dip 25°-70° N. on the south limb and southward at somewhat lower angles on the north limb.

In the Michigan Copper district, on the south limb of the basin, the lava series consists of at least 200 flows, with a few beds of rhyolite conglomerate between some pairs of flows. Many of the stratigraphic units (flows and groups of flows) are remarkable for their persistence and uniform thickness (fig. 1).

VOLUME OF LAVA IN INDIVIDUAL FLOWS

The thickest flow in the Michigan Copper district is the Greenstone flow (Cornwall, 1951, p. 164-166), which has a maximum thickness of more than 1,400 feet and is demonstrably more than 1,000 feet thick for a distance along the strike of 27 miles. It averages more than 200 feet thick for an additional 20 miles. This flow also crops out on Isle Royale (Lane, 1898, p. 99-102), across the basin 50 miles away, where it averages 300 or more feet thick for a strike length of over 42 miles. Ignoring the probability that the thickest part of the flow is beneath the lake, the volume of lava in this flow is more than 200 cubic miles by very conservative estimate.

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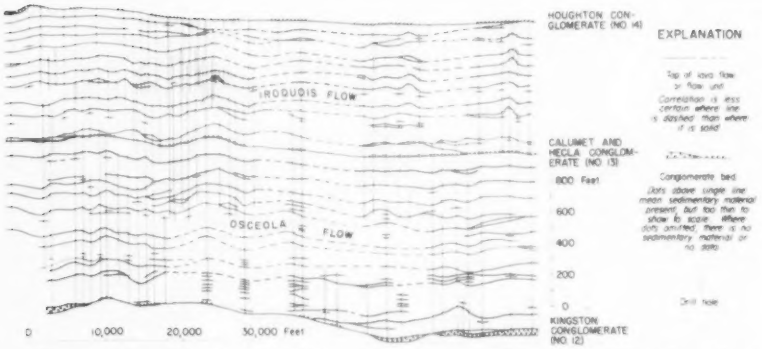
HOUGHTON CONGLOMERATE (NO. 14)
IROQUOIS FLOW
OSCEOLA FLOW
CALUMET AND HECLA CONGLOMERATE (NO. 13)
800 Feet
600
400
200
0
KINGSTON CONGLOMERATE (NO. 12)
0
10,000
20,000
30,000 Feet
EXPLANATION
Top of lava flow or flow unit
Correlation is less certain where line is dashed than where it is solid
Conglomerate bed
Dots above single line mean sedimentary material present, but too thin to show to scale. Where dots omitted, there is no sedimentary material or no data
Drill hole

Fig. 1. Longitudinal stratigraphic section of a group of Keweenawan lava flows. From left to right, section extends 4-15 miles northeast from Calumet.

Another flow, the Scales Creek ophite (White, Cornwall, and Swanson, 1953), is almost certainly persistent for a strike length of more than 55 miles, and its thickness, which is relatively constant, averages about 200 feet. If its areal extent even approaches the square of its strike length, the volume of lava is of the order of 100 cubic miles.

In the Michigan Copper district as a whole, 37 percent of the total volume of lava occurs in flows more than 100 feet thick, and 12 percent in flows more than 200 feet thick. A number of flows of this general size appear to be persistent with relatively constant thickness for distances of at least 25 to 30 miles, although one cannot normally prove this persistence for each flow individually because of the weaknesses inherent in correlating from drill hole to drill hole (fig. 1). But a flow only 100 feet thick covering an area 25 miles square has a volume of more than 10 cubic miles, and there are many tens, if not hundreds, of such flows in the Lake Superior basin as a whole.

That these volumes are not excessive for plateau lava flows may, perhaps, be shown best by an illustration from the Columbia Plateau. Waters (1955, p. 708) describes a flow 360-430 feet thick that can be traced 120 miles north-south and 50 miles east-west; the volume of lava in this flow may be conservatively estimated to be of the order of 300-400 cubic miles.

SLOPES

Although it is, of course, impossible to measure directly the original slopes of the tilted Keweenawan lava flows of the Lake Superior region, these slopes must have been very gentle—probably no more than 10 or 20 feet per mile—to satisfy certain limiting conditions that can be measured. Two factors that help define these limiting conditions are the fact that stratigraphic units thicken towards the center of the basin, and the evidence that the direction of slope was periodically reversed.

Thickening towards the center of the Lake Superior basin.—Thickening of stratigraphic units toward the center of the Lake Superior basin was conclusively demonstrated some years ago by measurements in the mine workings

of the deep Calumet and Hecla mine (Butler, and others, 1929, pl. 20). Although the measurements of rate and direction of thickening that can be made at this mine may long remain the best in the entire Lake Superior region, they can be supplemented by others of somewhat poorer quality over the much larger area in which there has been extensive drilling and underground exploration. All these measurements point to the same general conclusion: the rate of thinning in the Michigan Copper district, if persistent to the south, is such that the middle Keweenawan lava series must wedge out completely within 15 to 20 miles south of the present line of outcrop of the lava series. The picture of a basin subsiding as it was filled seems established beyond any reasonable doubt.

Evidence for periodic reversals of slope.—Bent pipe amygdules at the base of lava flows are fairly common in the Michigan Copper district, and show that the prevailing direction of lava movement was towards the margin of the Lake Superior basin (Butler, and others, 1929, p. 26). Conglomerate and sandstone beds between some of the lava flows, on the other hand, seem to have been deposited by streams flowing in the opposite direction (White, 1952, 1957). The evidence is based primarily on imbrication of pebbles in three different conglomerate beds, and on scattered observations of cross-lamination. In addition, the thick (3,000-5,000 feet) conglomerate formation above the lavas, which has the same general suite of pebbles as the interflow conglomerates, was definitely deposited by streams flowing toward the basin, as shown by imbrication, changes of facies, local abundance of pebbles of pre-Keweenawan rocks, and more than 250 measurements of foreset lamination. Although four different authors have briefly mentioned cross-bedding that suggests stream flow toward the margins of the basin (Hotchkiss, 1923, p. 672; Butler, and others, 1929, p. 26; Aldrich, 1929, p. 111; Sandberg, 1938, p. 820), these exceptions to a more general trend of drainage toward the basin can probably be explained as local irregularities due to meandering or diversion of streams by lava.

The evidence for basinward flow of streams, contrasted with evidence that the lavas flowed toward the margins of the basin, shows that there were, at times, reversals of the prevailing slope over large areas. These reversals permit one to set rather stringent limits to the original surface slopes of the lava flows.

Slope of Keweenawan lava flows.—The writer has suggested elsewhere (White, 1957, p. 6) that an imperfect balance between downwarping of the Lake Superior basin and filling of the basin by lava may have been responsible for the apparent reversals of slope. According to this interpretation, the lava was horizontal or sloped gently toward the margins of the basin as long as filling kept pace with downwarping. When extrusion was interrupted, however, downwarping produced basinward slopes over small to large areas. If this mechanism is correct, the angle between a conglomerate bed and the top of the immediately overlying flow is equal to the arithmetic sum of the initial dips of conglomerate bed and lava flow. At the time when deposition of a conglomerate bed was interrupted by renewed outpouring of lava, the conglomerate bed still had its initial dip, and the lava surface presumably had a small, but unknown slope in the opposite direction from that of the conglomerate. We could determine this unknown slope of the lava surface by subtraction if we

could (1) measure the angle between the top and bottom of a number of flows that overlie conglomerate beds, and (2) determine the approximate initial dip of the conglomerate beds.

The angle between the top and bottom of flows immediately overlying conglomerate beds can be measured in sections more or less normal to the axis of the Lake Superior basin at several places in the Michigan Copper district (table 1), although the data available are not as numerous or reliable as one might wish. Insofar as possible, the effect of local irregularities has been reduced by averaging a number of measurements, or, under still more favorable conditions, by making an isopach map of the flow and determining the angle from the spacing of contours.

TABLE 1

Angle between top and bottom of flows overlying conglomerate beds

A	B	C	D	E
Hancock	55	200	4	Poor
Allouez	85	230	Isopach map	Good
Calumet and Hecla	15	70	60	Fair
Kingston	80	135	Isopach map	Good
Wolverine (sandstone)	0-20	195	Isopach map	Excellent
St. Louis	55	115-570	6	Poor
Baltic	40	200	Isopach map	Excellent

Column A. Name of underlying conglomerate bed (see White, Cornwall, and Swanson, 1953; or Butler, and others, 1929, pl. 15).

Column B. Angle between top and bottom of flow (ft/mi).

Column C. Average thickness of flow (ft) in area of measurement.

Column D. Method of measuring: a numeral gives number of drill holes involved where measurement was between pairs or groups of deep and shallow drill holes, respectively.

Column E. Relative reliability of measurement.

The angle between top and bottom of the flows listed in table 1 ranges from 0 to 85 feet per mile. The Wolverine sandstone is a fine-grained sandstone, and may well have been essentially flat when deposited. The surface of the flow above it, therefore, probably had initial dips ranging from 0 to 20 feet per mile. From all the other angles, however, one must subtract an unknown amount for the initial dip of conglomerate beds in order to obtain the initial dip of the flows. According to the data of the table, this intial dip of the conglomerate beds was no more than 85 feet per mile (about 1 degree), and is therefore much less than the 3 degrees once suggested by the writer in an attempt to explain a peculiar characteristic of the imbrication in a conglomerate bed (White, 1952, p. 197). Although there are clearly too many unknowns to venture a calculation of the dip from hydrologic formulae, analogy with modern or near-modern conglomerates may be helpful.

The interflow conglomerate beds of the Michigan Copper district may be briefly characterized as follows: they are sheetlike bodies with lateral dimen-

sions that range from a few feet up to tens of miles, and thicknesses that range from a fraction of an inch up to several tens of feet (thicknesses in excess of 50 feet are uncommon); pebble conglomerate with a median particle size in the general range of 1-4 inches is very common, and boulders up to 3 feet across are found in some of the coarser grained beds. It might be pointed out, in passing, that the conglomerate beds (Allouez and Kingston) beneath the two flows with the largest angles in table 1 are among the most persistent and coarse grained of all the interflow conglomerate beds. The upper parts of conglomerate beds are not characteristically more silty or sandy than the lower parts. This suggests that the beds were deposited in some environment like an alluvial fan where aggrading streams sweep back and forth across the surface in very shallow channels, rather than in an environment of gentler slope such as (1) the flood plain of a river, where the gravel of the bed load in the channel is generally blanketed by finer grained material in the stream banks and on the surface of the flood plain (Hack, 1955, p. 33-35; Schlee, 1957, p. 1396-1398), or (2) the lower, very gently sloping part of a piedmont fan, which has the same general characteristics (Eckis, 1934, p. 87-88).

Gravel comparable in coarseness to the interflow conglomerate beds is generally associated, on fans, with slopes greater than 1 degree (about 92 feet per mile). On a fan 4 miles long in the Santa Catalina Mountains of Arizona, Blissenbach (1952) found slopes of 1 degree associated with maximum pebble sizes of only 12 inches (30 cm). On three smaller fans in the same region, the material was even finer grained at this slope. The gravel fans of the Elkton area, Virginia (King, 1950, p. 58-62, pl. 1), which are presumably representative of fans 2-3 miles long in a fairly humid region, have slopes that are almost everywhere more than 100 feet per mile.

On larger fans, Eckis (1934) found slopes as low as 90 feet per mile at a place where the maximum pebble size was about 30 inches; the median size in a measured sample (Sample G81, Eckis, 1934, Appendix) from approximately, if not exactly, this same locality, was about 1 inch (16-32 mm).

Although no gravels of recent origin have been deposited in an environment that is directly comparable to that of the interflow conglomerates of the Michigan Copper district, it seems safe to conclude from the observed relationships between slope and pebble size in modern fans—probably their nearest counterparts—that the slopes of the streams of middle Keweenawan time were at least several tens of feet per mile. Deducting such gradients from the angles of table 1 leaves little, if any, remainder for the initial dips of many of the flow tops. The flows, therefore, had very low gradients indeed, probably in the general range of 0 to 20 feet per mile. Judging from the persistence and uniform thickness of many flows and conglomerate beds (fig. 1), the areas over which such low gradients were characteristic were many tens, if not hundreds of square miles in extent. Very thick and widespread flows like the Greenstone flow probably had almost horizontal surfaces over areas of several thousand square miles.

VOLUME, SLOPES, AND MOTION OF THE FLOWS OF SHIELD VOLCANOES

The average volume above sea level of the 20 largest historic flank flows

of Mauna Loa and Kilauea (Wentworth and Macdonald, 1953, p. 31) is 0.037 cubic mile, and the volume of the largest flow is 0.11 cubic mile. If nine-tenths of the largest flo wis below sea level, its volume may exceed a cubic mile. Even the famous 1783 flow from the Laki fissure in Iceland is estimated to have a volume of only 3 cubic miles, plus another cubic mile expelled as ash, lapilli, and bombs (Thoroddsen, 1925, p. 60). Compared with historic flows from central or shield volcanoes, therefore, plateau lava flows like those of Michigan or the Columbia Plateau can demonstrably attain volumes an order of magnitude or two larger, and it is only one's imagination that sets an upper limit to their possible size.

The slopes of shield volcanoes are steep compared with the surfaces of plateau lavas. The slopes of Mauna Loa are mostly between 600 and 400 feet per mile, except close to sea level. Over most of the island of Hawaii, gradients are more than 200 feet per mile, and gradients of less than 100 feet per mile are of only local extent, as in saddles or along the crests of rift zones. Where the main 1783 flow from the Laki fissure in Iceland poured out on a coastal flat 30 miles and more below its source, an area of about 20 square miles has an over-all gradient as low as 10 feet per mile (gradients scaled from topographic maps), although over most of the coastal flat, 15-20 feet per mile is a more common gradient. The gradients between the coastal flat and the fissure itself are much steeper. Only locally, therefore, do the flows of shield volcanoes have slopes as gentle as those that seem to characterize flood basalts.

Motion of lava on shields necessarily ceases on relatively steep slopes. In aa flows, the center becomes increasingly crystalline and too viscous to move, and the whole flow stops. Pahoehoe flows of shields, during their extrusion, are normally covered by a continuous crust a short distance downstream from their vents; because the steep slopes of shields tend to make the flows both thin and narrow, this crust soon becomes anchored, whereupon movement of lava becomes increasingly confined to tunnels and the configuration of the flow as a whole is frozen.

MECHANISM OF BASALT FLOODING

The principal consequence of the great volume of lava floods is that on a relatively flat surface a flow may come approximately to rest while its interior, at least, is still molten, giving the top time to approach a hydrostatic level (Sandberg, 1938, p. 818). The near-level surface of such a flow provides the floor for a succeeding flow, making the upper and lower surfaces of succesive flows nearly parallel over large areas (fig. 1). As envisaged here, the formation of a basalt flood has its closest observed counterpart in the initial burst of lava from vents on shields. The great volume of lava and the gentle slope keep the flow thick, molten, and exceedingly wide, and do not permit the top to become anchored or tunnels to form. Motion stops primarily because the lava is ponded, either against an uphill slope or behind the dam of its own frozen margin. After eruption ceased, there would, of course, be some movement of lava from central to marginal parts of the flow, and some local advances of the front itself as long as the leveling process continued, but these movements would be minor compared with those of the initial flooding. The

crystallization of a large part of the interiors of flows would thus take place under esesntially static conditions.

CONCLUSION

The large volume of the flows and the probability that they crystallized in good part after movement ceased explain certain characteristics of the Keweenawan lavas that distinguish them from Hawaiian-type lavas. Among these may be listed (1) the fact that flows thin, rather than thicken, in the direction of flow; (2) the apparent absence of lava tunnels; (3) the differentiation shown by many flows (Broderick, 1935; Cornwall, 1951); (4) the complete lack of true tripartite aa flows (Wentworth and Macdonald, 1953, p. 59-62), and the abundance, instead, of flows that are fragmental only at the top (upper 18 percent of flow, on the average) and not at the bottom, and whose rubbly tops are composed of material that appears to be broken-up vesicular pahoehoe crust rather than spinose aa fragments. Columnar jointing, relatively rare in the Keweenawan lavas but exceedingly common in the Columbia River and other plateau lavas, might be added to this list. The first four characteristics, at least, seem to be independent of the composition of the Keweenanwan lavas, which mainly range from olivine basalt to andesite (Cornwall, 1951, p. 193-196). It is beyond the scope of this paper to inquire into the reasons why flood basalt flows like those of Michigan may have generally been about two orders of magnitude larger than those of shield volcanoes, but it seems entirely possible that this greater volume alone, rather than greater fluidity (Washington, 1922, p. 766, 803) due to differences in composition, volatile content, or temperature at the vent, is responsible for the major physical differences between the two types.

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TERRAIN INTELLIGENCE AND CURRENT MILITARY CONCEPTS*

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ABSTRACT. The development of nuclear weapons for tactical use has been followed by a concept of warfare based on small, mobile combat units equipped with helicopters, aircraft which can land and take off on short, sketchily prepared runways, and low ground-pressure vehicles which can move across country to supply a rapidly shifting force. This requires that terrain intelligence, dealing with the effects of the ground on military construction and maneuver, be made available to forward units that may have to act quickly on their own initiative. Help in this effort will come from new means of gathering information, such as radar, infrared photography, and television. Use of statistical methods will accelerate terrain analysis, especially at the strategic level. Analysis of landforms as they offer shelter from nuclear blast, and of soils which may emit neutron-induced gamma radiation, will assume an important place in terrain intelligence, as also will selection of safe routes for helicopters, tanks and supply vehicles; sites for rapid airfield construction and paratroop landings; and the suitability of soil for hasty excavation of shelters. The rapid pace of such warfare will require frequent preparation of situation maps showing areas where tree blow-down might slow troop movement or where nuclear contamination might be dangerous.

INTRODUCTION

Terrain intelligence, or terrain analysis, is "The process of interpreting a geographical area to determine the effect of the natural and man-made features on military operations. This includes the influence of weather and climate on those features." (Dictionary of U. S. Army Terms, 1953). This paper deals only with the analysis of natural features. Such analysis has, at least since the beginning of World War II, been widely referred to in the United States as military geology. Similarly in Germany, from World War I on, it was categorized as Wehrgeologie. In fact, however, as can be seen even in the above brief definition, the analysis of natural terrain for military purposes transcends even a liberal definition of geology. It includes meteorology, climatology, hydrology, soil science, botany, and forestry; and with modern developments in warfare geochemistry, geophysics and statistics have assumed great importance in this field. The knowledge of terrain derived from analysis by specialists of many sorts is combined in the intelligence report, a document, usually with a number of specially designed maps, which predicts terrain conditions of an area of military interest. The area of study is usually inaccessible, although terrain studies may also be made of maneuver areas or test sites for training or experimental purposes. Because terrain intelligence reports are designed for military users, those who write them can at present assume no technical knowledge on the part of the reader beyond that normally gained in military training (Whitmore, 1959). Furthermore, the user of such reports is usually pressed for time; therefore texts must be kept brief, and as much information as possible must be put on maps or in tabular texts which serve as expanded map explanations and enable the reader to look up terrain conditions at a given point of interest without reading through a running text.

The subjects included in terrain intelligence have increased in number as warfare has grown more complex, and this trend is, of course, continuing. Physiography was the first branch of earth science to be applied to military intelli-

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gence; this application was pioneered by the French military geographers (Marga, 1884-85, cited in Bryan, 1920). This example was followed by geologists assigned to the American Expeditionary Force in World War I (Bryan, 1920), who prepared terrain appreciation maps of parts of France for use at Corps level. The war of position, with its emphasis upon extensive networks of trenches with ancillary bunkers and tunnels, and the related problems of drainage and of water supply for large numbers of more or less stationary troops, led for the first time to a need for information on subsurface conditions (Brooks, 1920).

Between World Wars I and II, the Germans continued an interest in military geology and related applications of the natural sciences to terrain intelligence (Sonne, 1936; Wasmund, 1937; von Bülow, Kranz, and Sonne, 1938; Kranz, 1940). The results of their extensive studies in the field were summarized and applied to the warfare of the time in a series of 38 papers prepared for a course in military geology, held by the German army military geology organization at Heidelberg in 1940 (Kraus, et al., 1941).

In the United States, military geology was revived early in 1942 with the organization of the Military Geology Unit of the United States Geological Survey under the leadership of W. H. Bradley. Although Bradley's writing on the subject consists only of one abstract (Bradley, 1945), his influence upon the development of methods and formats for the preparation and presentation of terrain intelligence, and upon the use of the resulting reports by the U. S. Army, has been one of the major factors in the evolution of modern terrain intelligence. The Military Geology Unit, which by the end of the war had grown to a staff of about 130 scientists and supporting workers, prepared studies for many different military echelons dealing with cover, concealment, fields of fire, hasty fortifications, location of natural construction materials, foundation conditions, water supply, trafficability (suitability of the ground for cross-country movement by vehicles), airfield site selection, road construction and maintenance, and the effect of ground upon mine detectors. An account of the organization and methods of the Military Geology Unit, with examples of its work, has been prepared by Hunt (1950); additional information concerning the history of military geology is given by Whitmore (1954).

As modern terrain intelligence has evolved, all armies which use it have developed, more or less independently, the "combination method," a term coined by the Germans for the team approach, in which a number of specialists in different disciplines collaborate in the preparation of a report. Furthermore, three levels of intelligence have been generally accepted both by the users and the producers as a means of defining the use to which a study will be put, and thus its content, format, and degree of detail. The most generalized of these is *strategic intelligence*, prepared for the use of headquarters staffs down to field army level, and required for long-range planning. Strategic intelligence reports generally require map scales of 1:1,000,000 or smaller and deal with areas of sub-continental size. Next is *operational intelligence*, with map scales larger than 1:1,000,000 (usually 1:250,000 or larger), which are used in preparing plans for a specific military operation and for bringing intelligence information up to date as the operation progresses. Such intelligence reports are usually used

at field army headquarters or below (corps, division, battle group, regiment, or battalion). The third, and most detailed, category of terrain intelligence is *field intelligence*, a phrase coined here to include battlefield reconnaissance and surveillance, and also field consulting. Examples are field observation of the suitability of ground for cross-country movement, and of the detailed engineering geology characteristics of an area. Field consulting is a familiar function to the geologist, and has proved useful in the armies of the United States, Great Britain, Germany, and Russia. It involves such activities as location of water supplies, selection of quarry sites, search for sources of gravel, on-the-spot advice to engineer construction units as to road construction and maintenance, drainage, subgrade, base course and soil stabilization, and examination on the ground of sites that have been selected by operational intelligence as suitable for airfields or for underground installations.

All the major aspects of terrain intelligence as described above—team approach, concise format, specially designed maps, subject coverage, and levels of intelligence—will undoubtedly continue to characterize this activity in the future. But it is already apparent that modern concepts of warfare will alter each of these aspects in greater or less detail. New subjects of importance to military operations have appeared because of the advent of new weapons and vehicles. New means of gathering information force us to reopen the question of what information is useful to the military planner and how it should be presented; this is particularly important in view of the high level of technical training now required of officers in certain parts of the armed services. The resolution of these and many other problems will determine where and how the natural scientist can be of greatest use in terrain intelligence.

CURRENT MILITARY CONCEPTS

Factors influencing development of concepts.—The means of waging war determine how war is waged; and the three major "means" are weapons, transport, and communications. It is true now, as it has always been, that development of the various tools of war has not taken place at the same rate. Modern military thinking is, of course, more influenced by the availability of nuclear weapons than by any other consideration. In the past, innovations in weapons development have favored either the offense or the defense, thus focussing military thought on the development of ways of neutralizing the newly-gained advantage of the enemy. In the case of nuclear weapons the reaction of military thinkers has had to be more complex: not only is the obvious matter of defense and survival against enemy bombardment to be considered; there is the equally knotty problem of the use of the weapon to the tactical advantage of the user. The mission of land warfare is to neutralize the enemy's forces and to secure and hold his territory. In pursuing this objective by the use of nuclear weapons, the offensive commander must be aware of the troop safety factor, i.e., that there is a considerable possibility of damaging his own troops with his own weapons. The three effects of nuclear bombardment are blast, heat, and radiation (both immediate and delayed). Blast and heat effects can injure friendly troops if they are not warned of an imminent bombardment, or if there is an error in bomb delivery. This situation is, in a greatly magnified way, like the problems

of large-scale conventional artillery bombardments in World War II, when errors of placement sometimes put friendly troops under fire.

But, in terms of troop safety, it is the radiation effects which pose a new and difficult problem for the offensive commander. He can interdict an area from the enemy by the use of atomic weapons, but how can he occupy or cross it? Efforts to answer this question must have great influence on the development of future tactical doctrine and hence on the development of terrain intelligence.

Turning from the offensive to the defensive viewpoint, the fundamental premise in planning for nuclear warfare must be that the smallest possible target will be presented to the enemy. This can be achieved in one of two ways: either by going underground or by achievement of far greater mobility than troops have had before. The consideration of underground installations for defense is different in degree rather than in kind from similar defense against conventional weapons. The matter of mobility, on the other hand, forces radical changes in our concepts of tactics, and both military theory and specific vehicular development have been in recent years greatly influenced by this factor.

It is generally agreed (Parham, 1957) that, in two-sided atomic war, it will be necessary to separate groups of soldiers so that only a small number will be vulnerable to atomic bombing except possibly during a short massing period before an attack. This means that there will be gaps between fighting groups and thus an end to the classical concept of the continuous line facing the enemy. The existence of separate units of this sort means a great increase in the periphery to be guarded, in addition to which the gaps between fighting units must be kept under surveillance. This in turn means that each unit, after being well briefed in advance on its mission and objective, must be semi-autonomous within the limits of its mission. A great deal of intelligence information must be given it before action starts, but it must also be capable of collecting detailed information about the battlefield as it goes along. The unit must be able to move rapidly from place to place, and it must be supplied from depots in the rear. These requirements have led to the development of new vehicles, both ground and air. Troop-carrying helicopters have been designed, as well as an experimental series of turbo-rotor or "zero ground pressure" vehicles such as the individual flying platform and the "aerial jeep." For cross-country movement, especially of supplies, attention is being given to a class of vehicles knows as Goers, with large wheels and low-pressure pneumatic tires similar to the large earth-moving equipment now in wide use (McKee and Rigg, 1958). Air supply and movement of troops from place to place on the battlefield may also be achieved by the use of aircraft capable of vertical take-off and landing, and also of aircraft which can land and take off on short runways which have been prepared only to the extent of removing major obstacles and grading and filling the extremely rough spots.

Planning a military force which will make optimum use of these new developments both in weapons and mobility is further complicated if the planning is aimed at developing a force that has dual capability, i.e., that can fight either a nuclear or a non-nuclear war (Lemnitzer, 1959). Whether this is possible or advisable is a matter of disagreement among military authorities.

Army organization.—Since terrain intelligence is primarily concerned with land warfare, the organization and intelligence requirements of the U. S. Army, as influenced by the new developments discussed above, will be considered here, together with certain Air Force developments that relate to land warfare. By extension these considerations will also apply to amphibious operations of the Marine Corps and Navy and, on a broader scale, to strategic operations of the Air Force.

To meet the requirements of nuclear warfare, with the attendant necessity for quickly placing a striking force anywhere in the world, the U. S. Army has established a new type of unit, the Strategic Army Corps (Clarke, 1958, Westmoreland, 1958). This consists of two airborne and two infantry divisions with combat and service support elements such as an armored cavalry regiment, corps artillery, engineers, transportation, and quartermaster troops. The major elements of the Corps are ready for deployment overseas by air on a few hours' notice; the remainder could be deployed by air and surface lift shortly thereafter.

Each of the airborne divisions of the Corps consists of five battle groups, and each of these consists of five rifle companies, a headquarters company, and a heavy-mortar battery (Gray, 1958). The battle group, consisting of 1,584 officers and men, is the fundamental unit of mobile warfare; and the five-fold structure of division and battle group is the Pentomic concept which now governs the battle organization of the U. S. Army. A test of the Strategic Army Corps organization was made in May 1958 (Westmoreland, 1958). The 101st Airborne Division, a unit of the Strategic Army Corps stationed at Fort Campbell, Kentucky, received an alert at 2:00 P.M. on May 13. By daybreak on May 14 a task force, consisting of 560 men, 59 vehicles, one helicopter, and ammunition and rations for four days, was in the field in Puerto Rico, 1,600 miles from its base. For operations of this sort, initiated on a few hours' notice but likely to develop into campaigns of length similar to those of World War II, it is necessary to prepare terrain intelligence studies well in advance, in such a form that they can be easily carried and quickly used; and the initial studies must be supplemented by terrain estimates made by specialists who move to the operational area with the combat troops.

NEW DEVELOPMENTS IN THE APPLICATION OF THE NATURAL SCIENCES TO TERRAIN INTELLIGENCE

New means of gathering information.—A number of sensing devices, such as radar, infra-red photography, and television, have been developed to the point where they can be used in gathering terrain information. They can be used by troops in forward positions on the ground or by manned or drone aircraft flying over enemy-occupied territory, which can then transmit impulses (or, in the case of television, pictures) either to advanced command posts or to intelligence centers in the rear. In the case of radar, for instance, Feder (1959) emphasizes its ability to record terrain characteristics in all weather and to penetrate vegetation. He points out that "some X-through K_a-band reradiation data show radar's ability to interpret surface textures, moisture content, and snow

cover for trafficabi"ty purposes and texture and composition for civil engineering and geologic purposes." Terrain images received over radar are not clear enough to serve as the sole source of information on an area, but they can be a valuable adjunct to previously prepared terrain intelligence reports based on compilation from geologic, soil, and vegetation literature and from aerial photographs.

Another device for gathering supplementary terrain data is the portable reflectance spectrophotometer developed by the U. S. Army Engineer Research and Development Laboratories (Dwornik, et al., 1959). Spectral-reflectance data, which can be used in conjunction with photointerpretation, appear to be particularly useful in detecting areas where the soil has been disturbed.

Besides the new sensing devices, terrain intelligence will benefit from improved quality of aerial photographs and from increased speed of film processing and delivery.

Despite the many ways of gathering information at a distance, on-the-ground reconnaissance will still be necessary to determine the details of terrain configuration. Several instruments have been developed to measure soil bearing strength in the field to determine suitability for cross-country movement. Probably the most widely used of these is the cone penetrometer, described by Knight (1956). This is a rod tipped by a cone the area of whose base is one square inch, and, at the upper end, a flexible ring with a dial which records the pressure necessary to push the cone into the ground. The penetrometer measures the strength of the ground against vertical loads. It can be supplemented by a shearmeter, a ring-shaped load plate fitted with cleats. "This plate is applied to the ground with a constant vertical loading and provides a measure of the strength of the ground against horizontal thrusts by producing a record of torque *vs.* angle as the plate is rotated." (Sattinger, 1959). Such devices, while they probably could not be widely used under combat conditions, would be valuable in spot-checking existing intelligence, especially in areas of medium-grained soils whose reaction to precipitation might be difficult to predict with sufficient accuracy.

Quantitative terrain studies.—Terrain analyses have, since their inception, been for the most part qualitatively descriptive. Slopes are described as steep or gentle, ground as suitable or unsuitable for movement by tanks or as poorly drained or well-drained. Most often this description is presented in tabular form; examples are given by Hunt (1950). The exceptions to this rule have generally been in the form of stating limiting values: as, for instance, delineating areas with slopes too steep to be negotiated by a tank, or mapping areas of vegetation on the basis of whether or not the trees are far enough apart for a tank to pass between them.

As in most other branches of geology, there has been in military geology and its related sciences an increased effort at quantification in the last few years. The cone penetrometer is an example of this; it is used in an attempt to express the bearing strength of soil on a numerical basis. Besides trying to devise instruments specifically for military use, geologists and soil scientists have adapted the tests and measurements of civil engineering to reports for military engineers involved in the siting and construction of military bases, including

buildings, airfields, artillery and missile emplacements, roads, and underground installations. Nicol, Flint and Saplis (1957), in one of a series of military geology reports on islands of the western Pacific area that have been prepared by the United States Geological Survey, have presented the following numerical test results, in tabular form, for each of the units presented on their engineering geology map: mechanical analyses; physical test constants of fraction passing No. 40; specific gravity; moisture-density relation (undisturbed: field moisture; disturbed: optimum moisture; undisturbed: field density, lb./cu. ft.; and disturbed: optimum density, lb./cu. ft.); consolidation test with 0.341 tons/sq. ft. (initial void ratio and void ratio after 24 hours); California Bearing Ratio on water-soaked samples (% C. B. R. with 10 lb. surcharge and % swell during soaking); and direct shear test with 3,000 lb./sq. ft. vertical load. The figures given result from standard engineering tests of samples collected in the course of mapping the island. These samples were, in the judgment of the field geologist or soil scientist, typical of the map unit within which they occurred. The figures given are chiefly valuable in indicating order of magnitude of expectable conditions, and, especially, in comparing the worth of one site with that of another. It is understood that, before actual construction is begun, a detailed on-site survey must be made. This type of report has proved very useful to military planners.

Statistics as applied to landforms is a promising method that may be useful in future terrain analysis. The work of Strahler (1954, 1956; Strahler and Koons, 1959) and others in the field of quantitative geomorphology has been devoted to quantifying various landform characteristics in such a way that numerical values can be used to make comparisons between regions or to relate two or more categories of values to each other—as, for instance, slope related to the lithology of outcropping rocks. The application of statistical techniques to terrain analysis is difficult because of the number of factors involved. The preparation of a cross-country movement map for tanks is an example. The ability of a tank to move across country is governed by the following characteristics of the terrain: (a) slope; (b) soil: well or poorly drained, slippery or not; (c) state of the ground: frozen (to what depth?), snowcovered, flooded, saturated, or dry; (d) vegetation: presence and type of undergrush, size and spacing of trees, and amount and density of blow-down; (e) roughness (micro-relief); (f) distribution, size, and depth of water bodies. A particularly difficult aspect of this problem is that the resulting intelligence map is a simple one, classifying the ground in terms of a tank's ability to move (a) at any time, (b) at no time, or (c) seasonally (Whitmore, 1950, p. 644). This means that quantitative expression of landform characteristics will be a tool used in preparation of the intelligence report but will not appear in the report itself. This is in contrast to the presentation of the physical properties of soil and rock for construction purposes. The reason for the difference is that the construction engineer is accustomed to using numerically expressed properties in planning his operations, whereas the tactician, both in developing and executing his battle plan, must have his data reduced so they can be quickly absorbed and combined with the many other factors that require consideration in planning. A special-purpose

map, accompanied by a table or brief text, is the best way of doing this. It appears now that the major contribution of quantitative terrain studies to terrain intelligence lies in the possibility of quicker and more accurate delineation on maps of such important factors as slope classes.

Interpretation of an inaccessible area for military purposes requires intensive study of topographic, geologic, soil, and vegetation maps as well as of air photographs. The finished intelligence map results from a two-step process. First, the basic data are compiled. An example of this is the slope map. Then the various types of data are combined to form the interpretive map—one, for instance, dealing with cross-country movement. One of the most important aspects both of the basic-data map and of the interpretive map is the drawing of boundaries significant to military operations: an example is the line on the map representing the break between areas which a tank can negotiate and those which it cannot. Such a line on the interpretive map is usually derived from a consideration of the inter-action of two or more factors. For instance, the maximum slope that a tank con negotiate is one of the characteristics of the vehicle that is included in its design. This slope as presented in the official performance characteristics refers to the pitch of a straight, smooth ramp that the tank can climb. In the field, however, the maximum negotiable slope for the same tank will practically always be less than the maximum established by the manufacturers' test because of roughness of ground, slipperiness, and the presence of vegetation. A tank operating on a slope which approaches its maximum will be stopped by a tree or a fallen log which would present no obstacle on level ground.

Statistical methods can be applied to the preparation of basic data maps, each of which is based on only one, or at most a few, types of measurement. Even when one is making measurements on maps rather than in the field, time is a limiting factor; and the use of statistical sampling methods, analytical treatments such as frequency-distribution analysis, and curve-fitting and testing techniques (Strahler, 1954) may serve to accelerate the work of data compilation for intelligence purposes, particularly since machine methods may be used to handle statistical data. Allowance must, of course, be made for the plotting error of the base map; aerial photographs will be useful in determining this.

At present it appears that statistical analysis may be especially useful in strategic intelligence, where the area of a study is measured in thousands of square miles and the map scale is smaller than 1:1,000,000. Terrain analysis at such scales is made for the purpose of showing the distribution of mountains, foothills, plains, valley systems, and extensive poorly drained areas, and of pointing out their significance as barriers or corridors, or as suitable or unsuitable areas for military construction. A terrain unit on such a map may be several hundred miles long; and although it is predominantly of one landform type—foothills with maximum relief of 500 feet, for instance—it will naturally include within its extent other landform types which are too small to show at the required map scale. Statistical treatment may provide a rapid means of determining the dominant terrain type in such an area, together with a definition of its borders which is concomitant in accuracy with the map scale, and also may

furnish a good estimate of the nature and extent of subsidiary terrain types contained within the map unit.

In contrast to the strategic map, the large-scale tactical map, made to be used by troop commanders in the field, is probably not susceptible of statistical treatment. It attempts to show the exact conditions at a given spot rather than, as in the strategic map, the probability that a given condition exists there. Therefore direct compilation by skilled earth scientists, time-consuming as it may be, appears necessary in this case.

TERRAIN INTELLIGENCE SUBJECTS AFFECTED BY NEW CONCEPTS AND METHODS OF WARFARE

Terrain appreciation.—This term is used here for the analysis of the land surface as it affects maneuver. This includes cover; concealment; location of defiles, corridors, and barriers; suitability for hasty excavation (slit trenches, bunkers); and ease of observation. Cross-country movement for vehicles, because of the numerous special problems that it entails, is considered separately.

In nuclear warfare the troop safety factor will necessitate great accuracy in terrain intelligence concerning the area to be occupied by our troops. Troops can be protected from nuclear effects by being in the lee of a topographic barrier and by being in shelters, even hasty ones like slit trenches. Prominent terrain features will protect troops from thermal radiation, but the matter of protection from air blast effects is not as simple as it may appear. Studies at Hiroshima showed that some structures escaped destruction because there was a hill between them and the point of detonation of the bomb, but detailed analysis has demonstrated that blast waves can easily diffract around apparent obstacles (Glasstone, 1957, p. 36). These phenomena have two results as far as intelligence is concerned: first, a commander must, before making the decision to use a nuclear weapon, know the disposition of his own troops with respect to sheltering terrain; and second, the commanders of combat units must, before entering an area, have maps showing where they can dispose their troops for greatest safety, where the soil is suitable for digging shelters, and the location of existing underground shelter such as caves and mines. If we recall the mobility concept stated above, it becomes plain that battle groups and lower organizations will need to carry with them more detailed intelligence information than has been required in the past.

In addition to danger from blast and heat, troops in nuclear war will be endangered by fall-out and residual radiation. Fall-out patterns depend on winds and upon the shielding effects and eddies caused by landforms. The major protection against fallout from enemy bombardment will be the placing of troops in hastily dug covered shelters (again this emphasizes the fact that troops, no matter how often they move, will dig in at every opportunity). In planning for the safety of his own troops, the commander who contemplates using a nuclear weapon must be well briefed by his meteorologist.

One of the most dangerous types of residual radiation for troops in the field is neutron-induced gamma radiation, which results from the capture by certain elements in the soil of neutrons liberated in the fission process. Among

these are sodium and manganese, which can constitute a serious hazard during the first few hours after a nuclear explosion (Glasstone, 1957, p. 398-399). In order to be prepared for this danger, troops should have information as to the distribution of soils containing critical concentrations of dangerous elements.

In addition to intelligence studies prepared before the beginning of an operation, this type of warfare will require that terrain intelligence staffs prepare a sort of situation map at intervals during the fighting. This tactical intelligence, which should be distributed to front-line troops in each sector, would report on such recent developments as tree blow-down and newly contaminated areas, the location of which would affect the movement of troops from one place to another.

If the time should arrive when helicopters and vertical takeoff and short aircraft are used in forward areas for rapid deployment of troops, another map will have to be added to the already considerable number which comprise terrain intelligence. This would be an air movement map, pointing out low-level flight routes protected by valleys, ridges and forests. This same sort of intelligence would be useful for turbo-rotor vehicles and individual rotor platforms.

Recent developments in line-of-sight communications, such as television, and in surveillance equipment, make it necessary to determine for the battlefield area where the obstacles lie and where clear transmission or viewing is possible. Something on the order of Barrellian profiles would be useful in this determination. Snell (1959) has devised a mathematical model of terrain "from which it is possible to predict the number of valleys which cannot be seen (protected valleys) at specified distances from the highest elevation within an area."

Cross-country movement.—In World War II, except for the North African campaign, tanks generally moved on the roads most of the time, leaving them only when maneuvering room was needed in meeting the enemy, or when the road was threatened by aircraft or artillery. In a more mobile war off-road movement would become more important.

During World War II the United States Geological Survey designed a trafficability (now called cross-country movement) map, scale 1:100,000, for the medium tank. An example of a legend for such a map is given by Whitmore (1950). The 1:100,000 map has since been improved and simplified by Frederick Betz, Jr., A. C. Orvedal, M. M. Elias and their associates, and appears to be appropriate for the mobile warfare that has been described above, although further study of recent concepts as they develop might lead some to conclude that, because of the size of areas covered in an operation, the 1:250,000 scale would be preferable.

In mobile war waged by small units, logistics will be very difficult. Ammunition and rations for a battle group cannot be moved by air, and it seems unlikely that they can always, or even usually, be moved by road. McKee and Rigg (1958) state:

> We have made tremendous strides in increasing firepower. We cannot increase logistical power the same way. Until this has been accomplished, the sparsely populated battle zone will remain a myth . . . Road-bound vehicles can neither disperse nor concentrate with the speed and flexibility required in modern war, and dispersion is vital.

The Army is attempting to solve this problem with the "Goer" vehicle, mentioned above, having large cargo capacity, large wheels, and low pressure tires. The ground pressure exerted by these vehicles is near enough to that of the medium tank so that, with slight modifications in legend, the cross-country movement map designed for the medium tank could be used for the Goer as well.

The basic principle of dispersion in the battlefield would probably mean more hasty river crossings, as opportunity presented itself, and perhaps the virtual end of large-scale crossings at a single spot following the establishment of bridgeheads. Intelligence for this type of operation would necessitate location of all possible crossing points, no matter how marginal.

Surface- and ground-water contamination.—The possibility of water contamination by nuclear radiation will require a greater dependence on ground water by troops in the field than has been the case in the past, when surface water was always used when available.

Airfield construction.—Two constrasting trends are present in the siting, design, and construction of airfields. One is the result of the building of larger and heavier bombers and transport aircraft; the other, which has been discussed, is the advent of so-called "short takeoff and landing" aircraft.

In the development of larger conventional airplanes, Holle (1956) reports an increase in gross weight from a maximum of 17,000 pounds for the B-17 bomber of 1942 to 500,000 pounds for the B-52 of 1956. In situations where the early B-17 required an 8-inch concrete slab, the B-52 needs 21 inches. Modern large jet aircraft also require much longer runways. This means that selection of sites for larger airfields—a task of the military geologist—will be more difficult, although the techniques used will not differ much from those of the past. It will still be necessary first to select large flat areas where the surrounding uplands offer no topographic hazards to aircraft landing or takeoff, and where the subgrade material has sufficient bearing strength.

The use of short takeoff and landing vehicles does not impose a rigorous requirement in terms of length of runway. The need for specific soils information, on the other hand, is increased because, if used in close support of combat troops, these airplanes must land on runways whose preparation has been limited to removal of obstacles such as boulders and trees, followed by rough grading. Seasonal variation in the bearing strength of natural soils must be known, and the site which has been selected should be checked on the ground with a cone penetrometer or similar device.

An example of the selection of unprepared sites for aircraft landings is Operation Groundhog, carried out in North Greenland beginning in 1957 by the Air Force Cambridge Research Center and the United States Geological Survey (Stoertz and Needleman, 1957; Davies, Needleman, and Klick, 1959). A number of landing areas were selected by photo-interpretation and subsequently studied on the ground. One of these, on the shore of Bronlunds Fjord, was selected for a landing test by a C-124 heavy cargo plane.

> The landing area itself is situated on an exceedingly level, dry lake plain or old marine embayment, which is about one-half mile wide and is roughly parallel to the shore of Bronlunds Fjord for a distance of three miles . . . the southeastern one-third forming a natural landing area about one mile long, composed of hard, firm clay. (Stoertz and Needleman, 1957, p. 16.)

The aircraft which made the successful test landing had a gross weight of 145,000 pounds and ground contact pressure of about 54 pounds per square inch. A ground run of only 1800 feet was required for landing, and 2000 feet for takeoff. Preparation of the strip was limited to tamping of one small soft area, clearing grass clumps, and filling several shallow gullies with tamped earth obtained nearby.

Other subjects.—Selection of sites for paratroop landings and supply drops will probably be of increasing importance, but, with the development of helicopters and versatile fixed-wing aircraft, the emphasis on glider landings has already declined. Amphibious operations will be more dispersed than in the past, so that intelligence emphasis will have to include beaches previously regarded as marginal or unusable (Michel, 1958). The siting of underground depots and other major installations will be an increasing intelligence requirement.

ORGANIZATION AND DISSEMINATION OF TERRAIN INTELLIGENCE

In the type of warfare described above, the scale and amount of strategic terrain intelligence required will be about the same as in the past, but with greater emphasis on sites for hastily constructed airfields, air drop zones, air heads, and areas where topography offers protection for troops and low-flying aircraft.

Operational or tactical (i.e. large scale) intelligence studies must be disseminated more widely among forward units than has been done heretofore. Also supplementary situation maps and reports, dealing with tree blow-down, contamination, state of the ground and similar subjects, must be prepared as often as the situation warrants. The possible rapid shift from offense to defense will place a burden on intelligence units, and will raise problems in the design of intelligence reports. For defense, areas must be selected that are suitable for strong points, barriers, alternate positions, and blocking positions (Parham, 1957). For offense, possible attack routes must be shown and information kept up-to-date as to which ones remain feasible as the battle progresses.

In order properly to use the sensing devices that have been described, trained terrain intelligence personnel should be assigned to regiment or battle group headquarters, where they could report directly to the commander. Their basic data would consist of reports prepared at higher echelons before the operation, but they would add to and modify these with their own observations plus late information received from larger intelligence centers at Division and Army level. In nuclear warfare, once action has begun, it will probably proceed at such a rapid pace that most terrain intelligence reporting will be in the form of oral briefings or, if reproduced for dissemination, will be restricted to the bare essentials. In other words, it will be necessary that detailed terrain intelligence reports be distributed in advance not only to major headquarters, but to forward units which may find it necessary to act quickly on their own initiative.

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GENERIC CHANGE IN TERTIARY FLORAS IN RELATION TO AGE

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ABSTRACT. The question of an orderly, systematic change in generic composition with age of Cenezoic floras is reevaluated on the basis of 48 well documented floras from the western United States. These data reinforce the conclusion that there is an orderly, progressive increase in modernity of fossil angiospermous floras with decreasing age. The changing ratio of native to exotic genera in fossil floras, when compared to the modern distribution of these same genera, is of such consistency, as a function of geologic time, that generic analysis of temperate zone floras preserved either as megafossils or microfossils provides in itself a reliable adjunct in the determination of age. The age-generic change is viewed as a reflection of broad regional and essentially uni-directional climatic change from late Eocene to Pliocene. Data on the physiognomic characteristics of fossil leaves is correlated with the age of fossil floras and change in generic composition.

INTRODUCTION

Several years ago one of us (Barghoorn, 1951) presented a statistical analysis of some 32 Cenozoic and upper Cretaceous floras from wide ranging geographic areas of the United States. This analysis dealt with proportional generic composition in relation to geologic age. Each flora was divided into three generic categories (1) those genera represented in the deposit which are still extant in the area (2) extant genera currently exotic to the geographic areas of the deposit and (3) unidentified, extinct or form genera. The taxonomic treatment by the respective authors was followed in the case of each flora and the age assignment accepted as proposed on the basis of the evidence presented. Some evaluation was made with respect to form genera, in arriving at a value for the total number of generic entities. All data were limited to the angiosperms; the gymnosperms and lower vascular plants being eliminated from consideration. When the percentage of genera within each of the three categories was plotted against geologic time a functional relationship could be discerned between generic composition and age. This relationship was much more clearly defined in those fossil assemblages which are still represented in the modern flora of the region, i.e. the native generic element. The relationship was less well defined with respect to the exotic element and rather diffuse with regard to the unidentified or form generic category. The trend showed that between the upper Cretaceous and the middle Eocene there is a progressive decline in the "native" generic element and from early Oligocene to late Pliocene an almost exponential increase in this element. Conversely the exotic generic element increased to a maximum in the late Oligocene and rapidly declined in the Miocene and Pliocene. The unidentified, extinct or form generic category was randomly scattered in the late Cretaceous and early Tertiary but progressively reduced in the Miocene and Pliocene, reaching zero in the later Pliocene. It was evident from the scatter of the plotted data in category 2 (identified, exotic genera) that various authors make generic assignment with more confidence than others. It should also be noted that the vast majority of the generic identifications of the floras examined were based on foliar organs.

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In a sense the generic analyses and resulting plotted age-curves are a graphic expression of a phenomenon long recognized in the study of Cenozoic floras, viz. the older the flora, the greater the deviation from the plant assemblages which feature the modern vegetation of an area. On the species level, this principle was very clearly demonstrated by Reid (1920) in studies of the Pliocene floras of western Europe. Percentage composition and change therein with reference to age was greatly extended by Szafer (1946, 1954) in his extensive studies of the Pliocene floras of central Europe. Szafer's work, again, is on the taxonomic level of the species, but he extended and elaborated the concept of statistical change with time by segregating the component species of the fossil assemblage, whenever botanically determinable, into floristic elements conforming to current patterns of geographic distribution. It is of interest to note, that the percentage of *species* now *native* to the region declines from 98 percent in the Pleistocene (interglacial) to less than five percent in the upper Miocene as shown in a series of fossil floras secured from north of the Alpine-Carpathian mountain arc (Szafer, 1946).

Axelrod (1950) has aptly expressed this generalized floristic change with time, although not in quantitative terms, as follows:

> The Piru Gorge flora is considered to be of Middle Pliocene age because most of its species, like those of Mt. Eden and other Middle Pliocene floras of California, are similar to trees and shrubs now found within a short distance of the fossil locality.

It is evident, however, that a functional relationship between floristic change and geologic time must be limited to the most recent geologic periods if attention is restricted to the species level of taxa in fossil assemblages. In fact, in the case of higher plants it may be questioned whether it is scientifically justifiable to attempt to compare extant species with fossil species which are older that the later or latest Pliocene, if not Pleistocene. Unlike populations of living organisms, among fossils it is the genus rather than the species which constitutes the more realistic taxonomic entity by which correlations in time and space may be attempted and portrayed. Moreover, in comparing geographically distinct fossil assemblages recorded variously as megafossils, in the form of leaves, fruits and seeds or wood, with those recorded as microfossils in the form of pollen, it is patent that species comparisons as such are impossible. However, if isolated and detached organs are reduced, so to speak, to their botanical common denominator, the genus, then comparison and correlation may be attempted between populations of genera in stratigraphically and geographically separated fossil assemblages.

If a consistent statistical relationship can be established between geologic age and the percentage of genera in fossil floras *which yet survive* in the respective geographic regions it follows that the generic analysis of fossil floras provides a relatively simple and practical adjunct to age assignment and stratigraphy. In order to validate such a statistical correlation it is necessary that the relationship be established over a large unit of time and over a wide latitudinal and longitudinal range. It is also necessary, for reasons defined above, that correlations be limited to generic categories and should not incorporate the more transitory concept of fossil species. These criteria seem to have been

met satisfactorily in the previous study (Barghoorn, 1951) with respect to Cenozoic floras. The significance of the limited amount of data relating to the late Cretaceous is open to question owing to the uncertain taxonomic status of most Cretaceous genera of angiosperms. Moreover, most of the Cretaceous floras which were included have not been revised in recent years and remain in an unsatisfactory status.

Strong objection to what is termed the age-curve analysis of angiosperms has been expressed by Axelrod (1957) in a comprehensive criticism of the entire procedure of age asignment by generic analysis. Because much additional and reliable information has accumulated since 1951 regarding the botanical composition of many Cenozoic floras it is of considerable interest to examine this criticism, to re-test the validity of the earlier data and to refine the methodology by use of other paleoecological correlatives. Many of the criticisms expressed by Axelrod we feel reflect misapprehension of the broad regional generalizations implied. Certainly, if attention had been restricted to category (1) of the fossil floras, e.g. the "native" generic element it would mitigate criticism of the relative proportions of categories 2 and 3, for the percentage of category 1 will not change regardless of how much shift is made between categories 2 and 3. As for dual age possibilities in the case of the Brandon lignite, the botanical composition of this flora, well recorded as fruits, seeds and pollen, precludes any possibility of a Cretaceous age. On the other hand inclusion of Cretaceous floras presents a point of probably valid objection. In the past decade the great advances in palynological study of angiosperm fossils demonstrate how very few genera of existing angiosperms can be identified from Cretaceous sediments (Steeves, 1959). In fact, it is doubtful whether but a small fraction of Cretaceous genera can be assigned with any certainty to extant genera, and the accuracy of assignment is probably least among floras represented solely by foliar organs, without associated reproductive structures.

CHANGE IN GENERIC COMPOSITION WITH TIME OF CENOZOIC FLORAS OF THE WESTERN UNITED STATES

With the foregoing considerations in view the authors have confined this study both stratigraphically and geographically to the Cenozoic deposits of the United States west of the Rocky Mountain front, and have eliminated from consideration all floras which have not been revised or reworked within the past three decades. In order to provide a sounder basis for statistical analysis, all Pliocene floras of less than 10 genera and all Miocene or older floras with less than 15 genera have been excluded.

Because of the extraordinary thickness and completeness of the Cenozoic non-marine section in the western states a very large number of time-stratigraphic units are amenable to analysis and correlation. In this study a selected 48 floras have been considered. The sediments represent a time interval extending from early Eocene to the later Pliocene. More than half of the floras involved are of middle and later Tertiary time and are best represented in the Miocene sequence. Oligocene floras are smaller in number and a regrettably few earlier Tertiary (Eocene) floras are well documented.

In this study two rather than three categories of generic entities are

recognized, viz., "native" and "non native". The distinction between "exotic" and "unidentified or extinct" loses statistical validity if attention is focussed, as it is here, on the "native" element and its change with time. An extinct genus is obviously exotic to any region of the earth today and thus may be grouped with the exotic element for purposes of statistical treatment. The definition of a native genus is the same as proposed earlier (Barghoorn, 1951). A genus is considered native to a region even if it is locally absent because of orographic restriction in relatively recent geologic time. For example, although *Quercus* is not now native to eastern Oregon, the genus occurs in late Miocene floras both east and west of the Cascade range. It would not be justified to consider *Quercus* a native genus in the late Miocene localities west of the Cascades, while considering it exotic to localities east of the Cascades. In general the native genera have been defined on broad regional terms with reference to their known distribution as established by phytogeographic and floristic studies. Wherever data are available both microfossils and megafossils are included. It is interesting to note that in the case of floras known both as microfossils and megafossils (Faraday, Stinking Water, Blue Mountains, Mascall, Green River) the generic composition yields very similar ratios in each category, although the generic entities within each category do not botanically always coincide to a high degree. The often considerable degree of dissimilarity in generic lists drawn from microfossils in contrast to megafossils in the same units of a formation are due to the vagaries of sedimentary processes. It is evident that a truer picture of the total vegetation of a region can be obtained only by study of all categories of plant organs, if present, than by any one category (Barghoorn and Spackman, 1950; Traverse, 1955). Unfortunately most fossil floras do not provide evidence from a wide range of plant parts and hence are commonly little more than tokens of the actual source vegetation of the time and place.

The basic data are given in table 1, and the percentage of genera native to the region of each of the 48 fossil floras under consideration is plotted against time in figure 1. The time units designated on the abscissa are widely accepted subdivisions of the Cenozoic, but the actual spacing of the data within these units is in accordance with biostratigraphic zones based on the concatenation of evidence from faunal and other stratigraphic data. Insofar as possible these "zones" are based on the general principles of Oppel, whose work on detailed sections of the Jurassic is well outlined and elaborated by Arkell (1956). The "zones" thus deal with the definable ranges of particular species or genera. On the basis of biozones it is possible to make certain groupings of fauna or flora in the sequence of rocks in a given area. For example Latah-age floras (Miocene zone 2) can be recognized by the fact that this is the earliest occurrence in this area of forms such as *Quercus simulata*, *Fagus idahoensis* and several other species *and* that these rocks *also* contain the last known occurrence of such forms as *Exbucklandia oregonensis*. Subdivision of the Cenozoic of the northwestern United States into zones is most satisfactory from detailed sections of the Miocene and Pliocene, less so for the Eocene and earlier Oligocene.

Examination of figure 1 shows that the points lie along a curve, the trend of which shows a slow decline through the late Eocene and a consistent and

TABLE 1

Generic composition and leaf physiognomy of 48 Cenozoic floras from the western United States

Zone	Flora	Source of data	Location	# of genera	% native	# species	% non-entire
P-2	1. Alvord Creek	Axelrod, 1944	S.E. Oregon	14	93		
	2. Napa	Axelrod, 1950	Cen. Calif.	14	93	23	57
	3. Sonoma	Axelrod, 1944	Cen. Calif.	27	85	36	61
	4. Piru Gorge	Axelrod, 1950	S. Calif.	11	91		
P-1	5. Oakdale	Axelrod, 1944	C. Calif.	12	83	16	62
	6. Anaverde	Axelrod, 1950	S. Calif.	15	80		
	7. Mt. Eden	Axelrod, 1950	S. Calif.	25	83		
	8. Mulholland	Axelrod, 1944	C. Calif.	28	75	43	61
M-5	9. Dalles	Chaney, 1944	N. W. Oregon	12	67	14	78
	10. Troutdale	Chaney, 1944	N. W. Oregon	20	60	23	74
	11. Mint Canyon	Axelrod, 1950	S. Calif.	23	65		
	12. Middlegate	Axelrod, 1957	C. Nevada	19	63	37	65
	13. Table Mountain	Condit, 1944	C. Calif.	25	61	29	52
	14. Neroly	Condit, 1938	C. Calif.	18	57	53	67
	15. Remington Hill	Condit, 1944	C. Calif.	21	62	76	72
M-4	16. Faraday	unpublished data	N.W. Oregon				
	16.a. megafossils	unpublished data		25	56	27	78
	16.b. microfossils	unpublished data		27	52		
	17. Stinking Water	Chaney & Axelrod, 1959	S.E. Oregon				
	17.a. megafossils	Chaney & Axelrod, 1959		20	55	31	81
	17.b. microfossils	Chaney & Axelrod, 1959		18	56		
	18. Aldrich Station	Axelrod, 1957	C. Nevada	16	56	22	68
	19. Thorn Creek	Smith, 1941	S.W. Idaho	20	55	38	87
	20. Lower Idaho	Dorf, 1938	S.W. Idaho	21	57	25	74
	21. Blue Mountains	Chaney & Axelrod, 1959	N.E. Oregon				
	21.a. megafossils	Chaney & Axelrod, 1959		21	52	25	88
	21.b. microfossils	Chaney & Axelrod, 1959		22	50		

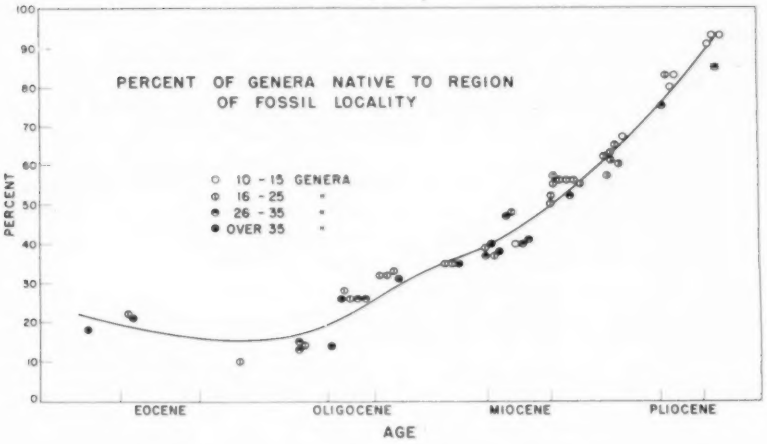
TABLE 1 (Continued)

Zone	Flora	Source of data	Location	# of genera	% native	# species	% non-entire
	22. Succor Creek	Brooks, 1938	S.E. Oregon	29	41	45	78
	23. Mascall	Chaney & Axelrod, 1959	N.E. Oregon				
M-3	23.a. megafossils			30	40	51	77
	23.b. microfossils			12	40		
	24. Upper Cedarville	La Motte, 1936	N.W. Nevada				
			N.E. Calif.	25	48	28	79
	25. Trout Creek	MacGinitie, 1933	S.E. Oregon	30	47	31	78
	26. Latah	Berry, 1931, Brown, 1937	N.E. Washington	55	38	65	76
	27. Grand Coulee	Berry, 1931	N.E. Washington	24	37	included in 26	
M-2	28. Tehachapi	Axelrod, 1939	S. Calif.	42	40		
	29. Russel Forests	Beck, 1945	S.W. Washington	27	37		
	30. Whitebird	Brown, 1937	N.W. Idaho	23	39	included in 26	
	31. Collawash	unpublished data	N.W. Oregon				
M-1	31.a. megafossils			94	35	11	75
	31.b. microfossils			20	35		
	32. Eagle Creek	Chaney, 1920	N.W. Oregon	20	35	33	76
	33. Bridge Creek	unpublished data	N.W. Oregon	50	31	53	75
O-4	34. Willamette	unpublished data	N.W. Oregon	24	33	25	72
	35. Rujada	Lakhanpal, 1958	S.W. Oregon	25	32	29	72
	36. Lyons	unpublished data	N.W. Oregon	25	32	27	77
	37. Weaverville	MacGinitie, 1937	N. Calif.	33	26	31	52
	38. Lower Mehama	unpublished data	N.W. Oregon	31	26	34	40
O-3	39. Princeton	unpublished data	S.W. Brit. Col.	22	26		
	40. Republic	Brown, 1937 & unpublished data	N.E. Washington	18	28	mega & micro-fossils combined	
	41. Florissant	MacGinitie, 1953	C. Colorado	73	26		
O-2	42. Goshen	Chaney & Sanborn, 1933	N.W. Oregon	36	14	49	45

TABLE 1 (Continued)

Zone	Flora	Source of data	Location	# of genera	% native	# species	% non-entire
	43. Veronia	unpublished data	N.W. Oregon	16	14		
O-1	44. Upper Clarno	Scott, 1954, & unpublished data	N.W. Oregon	31	13	(woods and seeds)	
	45. La Porte	Potbury, 1935	C. Calif.	33	15	36	36
E-5	46. Comstock	Sanborn, 1935	N.W. Oregon	21	10	27	33
E-4	(none well known)						
	47. Green River	Brown, 1934 Wodehouse, 1933	S.W. Wyoming				
E-3	47.a. megafossils			56	21		
	47.b. microfossils			18	22		
E-2	48. Chalk Bluffs	MacGinitie, 1941	C. Calif.	62	18	71	55

Fig. 1.



accelerating rise throughout the Oligocene, Miocene and Pliocene. This curve, depicting the "native" generic element and its numerical change, may perhaps best be regarded as a curve of climatic change, the rate of which increases in the later Cenozoic. It is most probable that the southward migration of more temperate forest types, replacing the older more tropical elements fundamentally influences the data. This phenomenon has been remarkably well documented in studies of the European Tertiary flora and its phytogeographic history (Reid and Chandler, 1933; Kirchheimer, 1957), and is likewise well supported by the smaller body of data available from eastern North America. Even the complicating factors of orographic change which affected the western third of North America far more than northwestern Europe and eastern America have not introduced major discordant factors in the generalized relationship between age and change in generic composition.

PALEOCLIMATOLOGIC EFFECTS AND EVIDENCE

It has been pointed out, in criticism of the age curve (Axelrod, 1957) that the thesis is botanically unsound because climatic change proceeds at different rates in different geographic regions. Except in lower latitudes of the true tropical regions this criticism does not seem to be tenable unless one is dealing with small time units and specialized regional differences such as insular endemism. The idea that present temperate zone "relict" floras have persisted in their present ranges over very long periods of time because climatic stress has not affected them is certainly not supported by any systematic body of factual evidence, nor is it consistent with any theory of planetary climatic change, such as the known facts of Pleistocene climatic history present. Indeed in the classic areas of "Tertiary relicts" in southeastern North America and southeastern Asia it is doubtful if many of these genera had entered the regions involved until the later Tertiary, and many are probably Pleistocene

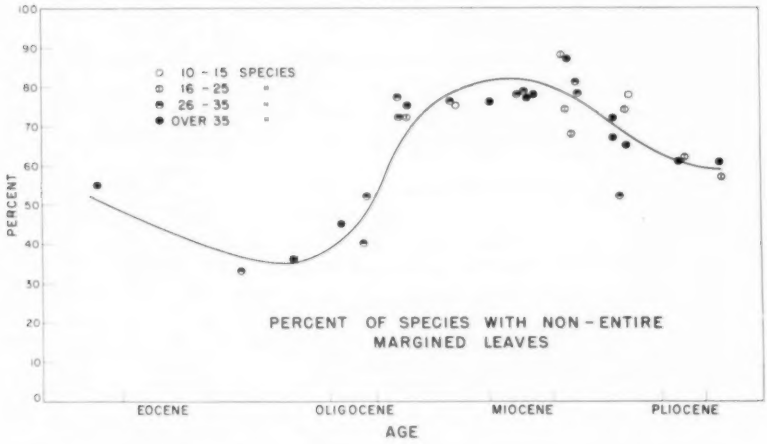
re-entrants. These are truly relict floras in the sense that they have achieved disjunction through extinction in intervening areas, not through persistence in geographic position *despite* climatic change. Our knowledge of the Cenozoic climatic history of the humid low latitude tropics is so meagre that it is not possible to relate it to the detailed picture which has been drawn for temperate, and chiefly cool temperate regions. It seems probable, however, that the tremendous thermal capacity of low latitude air masses and oceanic basins has buffered climatic changes to a degree which finds no counterpart in higher latitudes. The magnitude of climatic change within any time unit in the Cenozoic therefore would be less with decrease in latitude. The relatively small latitudinal range which we are dealing here (ca.32°N. to 46°N.), however, does not allow discrimination along these lines.

From a source quite independent of generic analyses, and in part devoid of paleobotanical comparisons with modern plants, another generalized concept of Cenozoic climate may be secured. This is derived from the physiognomic analysis of leaf form and structure in relation to climate and latitude in the extant flora. Data assembled from many thousands of species of tropical, subtropical, temperate and high latitude regions show that leaf margins and environment are consistently correlated (Bailey and Sinnott, 1915, 1916). Thus, leaves of woody plants (herbaceous plants were excluded) with *entire* margins tend to predominate in tropical, alpine and arctic regions and in physiologically arid environments. On the other hand, leaves (and leaflets) with *non-entire* margins are comparatively infrequent in such environments but are very numerous in moist temperate regions featured by cold winters and warm summers. Hence in the tropics and other regions of physiological aridity in terms of transpiration stress, the percentage of trees and shrubs with non-entire margins is low. The percentage increases in temperate and especially in humid temperate forest climates. If the climatic significance of leaf form, as demonstrated among living plants, be extrapolated to Cenozoic floras, it would provide an interesting paleoecological corollary to evidence of climatic change drawn from other sources. This has been proposed in paleoecological interpretations of Tertiary floras in several studies of the western Cenozoic (Chaney, 1927, Sanborn, 1933, Potbury, 1933, and MacGinitie, 1937) to augment other evidence of the climates in which various Tertiary floras existed.

Figure 2 presents the result of plotting the percentage of non-entire leaves against age. The data are derived from a selection of the major floras graphed in figure 1, and were secured from examination of published photographs of leaves and/or leaflets of named species. Many of the collections from which these floras were described are deposited in the collections of the Department of Paleontology, University of California, Berkeley, and have been re-examined by the first author.

It will be noted that the leaf-margin curve during the early Tertiary continues a consistent trend until the mid-Miocene, and then deviates sharply in a downward trend in the late Miocene and through the Pliocene. It is our view that this curve reflects a major change in vegetation induced by a fundamental change in climate affecting western North America. In the early Tertiary a humid mesophytic, in part subtropical forest, as evidenced by numerous genera,

Fig. 2.



is indicated by a high percentage of entire margined leaves. The subtropical elements are slowly replaced by more temperate types during Oligocene and early Miocene time. It is probable that the climate throughout much of our area during this interval of the Cenozoic was featured by mild winters and summer rainfall. The sharp reversal of the trend in the late Cenozoic seems most reasonably to record the onset of the Mediterranean type of climatic regime in coastal regions and the development of orographic aridity in the interior. It should be noted in connection with the leaf margin curve that the western interior major floras, the Green River and Florissant, are excluded so that the comparative aspects of the vegetational change are restricted to the more uniform physiographic and climatic province of the Pacific coastal states and the adjoining western interior basin.

In the leaf margin data, considerable scatter of the points among the Mio-Pliocene floras is interpreted as indicating increasing provincialism resulting from orogenic activity such as the rising of the Cascades. Hence contrasting local environments set in within the northwestern states and the rising percentage of entire margined leaves (decrease in non-entire) indicates encroaching aridity in the interior basins while the coastal climate tends to remain more stable and moist. It should be emphasized in this connection that drier environments of temperate regions are featured by a higher percentage of entire margined species among the woody flora than are more humid environments.

CONCLUSIONS

In evaluating the data here summarized we wish to emphasize that

1. A valid functional relationship exists in temperate latitudes between the age of a fossil flora and its generic composition. With time this relationship throughout much of the Cenozoic is featured by steadily increasing modernization on a broad regional scale.

2. There is no evidence that the rate of modernization, as indicated by generic change, varies greatly from region to region within the area considered in this study.
3. The generic composition of a fossil flora in terms of the statistical relationship between "native" and "exotic" genera is a valuable adjunct in age determination and stratigraphy. It must, of course, be regarded as merely one more line of evidence in geochronology and stratigraphy.
4. The dual relationship between age and the native vs. exotic proportion in fossil floras makes possible correlation in time (not necessarily floristically) between floras: a) widely separated in space, and b) known from dissimilar plant organs. Hence it is possible to correlate chronologically floras known only from leaves with those known only as pollen grains or fruits and seeds. This relationship is a useful adjunct in the palynological determination of age of Cenozoic floras known only from subsurface sampling.
5. Study of the statistical distribution of leaf form (entire margined vs. non-entire margined) in Cenozoic floras correlates with inferences drawn from generic analysis in showing progressive and broad regional climatic change.
6. There is no evidence from this study that the great climatic events characterizing the Pleistocene were of sudden origin. Rather the climatic deterioration featuring the Plio-Pleistocene transition came on gradually and was the culmination of events which began in the latter part of the early Cenozoic.

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BEDROCK PATTERNS AND STRIKE-SLIP FAULTING IN SOUTHWESTERN CALIFORNIA

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ABSTRACT. In southwestern California the pre-Turonian bedrock may be divided into 8 units, which may then be used to distinguish 3 rock belts. One belt comprises the glaucophane-bearing rocks near the shore, a second is characterized by quartz-rich graywackes, and the third, which is farthest northeast, contains limestones. The second and third belts include Mesozoic volcanics and extensive Cretaceous quartz plutonites. The bedrock mass has been disrupted by faults, some of which have been the loci of strike slips measurable in tens of miles. Much greater displacements have also been suggested. The simpler of these suggestions seem inconsistent with the distribution of the extrusive and intrusive Mesozoic igneous rocks. Perhaps on some faults two slip solutions are possible. In such cases tentative choices may be made by the use of the rule *Disjunctiones minimae, disjunctiones optimae.*

INTRODUCTION

The part of California considered in this paper is shown on the accompanying map (fig. 1). Los Angeles is in the west-central part of the area and San Diego near the southern margin. The map area covers a large and representative part of southwestern California but leaves out a block of land west of the north half of the map, divided politically between Ventura and Santa Barbara Counties.

In the map area the rocks are obviously divisible into two parts, a bedrock complex that is mostly crystalline and a superjacent series that is mostly sedimentary and in some places grades into the sediments now accumulating. A great unconformity separates the bedrock complex from the superjacent series.

In some places the unconformity separates rocks that differ only slightly in age. The upper Turonian ammonite genus *Subprionocyclus* has been found in the Santa Ana Mountains, southeast of Los Angeles, in beds some hundreds of feet above the base of the superjacent series. The lowest superjacent strata are non-fossiliferous conglomerates, in part red, that may also be Turonian. The bedrock, beneath the great unconformity, is rather certainly wholly pre-Turonian. In the Santa Ana Mountains, bedrock strata contain Upper Triassic marine fossils. Some 180-200 miles farther south, in Baja California, the youngest stratified bedrock unit, a clastic volcanic formation, contains *Douvilleiceras*, an Albian ammonite. These volcanics are intruded by quartz diorite. The quartz diorite is part of the plutonic assemblage that makes up more than half of the bedrock complex, in southwestern California as well as in northwestern Baja California. That is, the greater part of the bedrock mass may have solidified at plutonic depths in the brief interval between some part of Albian time and the beginning of the Turonian.

Geographically, the map area may be divided into four principal parts (cf. Hinds, 1952): two mountainous areas, one elevated flatland that contains isolated hills and mountains, and one lowland. The mountainous areas are the Peninsular Ranges, extending south from the latitude of Los Angeles, and the east-west trending Transverse Ranges north of Los Angeles. The two northernmost Peninsular Ranges are the Santa Ana Mountains at the west and the San Jacinto Mountains at the east. These two ranges are separated by low plateaus

sometimes called parts of the Perris Block. The principal Transverse Ranges in the area of figure 1 are the low Santa Monica Mountains at the west, which are mostly made up of the superjacent series, and the higher San Gabriel and San Bernardino Mountains farther east, which are mostly bedrock. The elevated flatland is the Mojave Desert, north of the Transverse Ranges. The lowland is an irregularly shaped area between the Transverse and Peninsular Ranges. It is narrow at the east and broad at the west. Here it will be called the South Coastal Basin, using that term in a geomorphic and structural, rather than a hydrographic, sense. The western and southern parts of the South Coastal Basin, including most of the city of Los Angeles, make up the Los Angeles Basin, in the center of which the bedrock complex has a sedimentary cover some 30,000 feet thick. San Diego is built mostly on a broad marine terrace that flanks a Peninsular Range.

Structurally the region is characterized by faults. Some of the larger faults are shown in figure 1. Even some of those not shown may have miles of strike-slip, dip-slip, or oblique-slip displacement. The San Gabriel, Santa Monica, and Santa Ana Mountains are cut into small blocks by mappable faults, some with straight, most with crooked traces. Faults seem to be relatively few in the western part of the Mojave Desert and in the northern part of the Perris Block.

In this paper the pre-Turonian bedrock is divided into 8 units. These units, which are the ones shown in figure 1, will be described briefly. Map and descriptions will then be used as bases for a final discussion that includes consideration of possible strike-slip movements of hundreds of miles on the San Andreas and other major faults of the region. The map units are not all age groups. For this and other reasons their order in the map explanation, and in the text, is only partially chronological.

ACKNOWLEDGEMENTS

Figure 1 was compiled from many sources, most of which will be noted in connection with the descriptions of map units. The most important sources were a published map by E. S. Larsen Jr. and unpublished maps by T. W. Dibblee Jr. Larsen's map (1948) covers the square degree between lat. 33° and 34° N. and long. 117° and 118° W., the northwestern part of the Peninsular Ranges. Dibblee's maps cover the part of the Mojave Desert shown in figure 1, the San Gabriel Mountains east of long. 118° W., and the San Bernardino Mountains west of long. 117° 15′ W.

During the preparation of this paper I discussed the rocks, structures, or age determinations with C. R. Allen, A. K. Baird, T. W. Dibblee Jr., P. L. Ehlig, R. H. Jahns, D. L. Jones, Robert McColl, T. H. McCulloh, Richard Merriam, J. F. Richmond, J. E. Schoellhamer, H. P. Schwarcz, J. S. Shelton, and R. F. Yerkes. Merriam and Yerkes loaned me rock specimens and thin sections. I have drawn freely on unpublished stratigraphic and structural discoveries by Ehlig and Schwarcz. Figure 1 was drafted by Toni Dana.

DESCRIPTION OF BEDROCK UNITS

Anorthosite and norite

Lithology and distribution.—The anorthosite-norite complex (Higgs, 1954; Miller, 1934) is a mass 15 miles long, in the western San Gabriel Moun-

tains. The anorthosite, which is 95 percent or more bluish-gray to white andesine, in crystals commonly half an inch or more across, grades into marginal, dark-gray andesine-hypersthene norite. Irregular bodies of apatite-ilmenite rock are present, especially in the transition zone. At the northeast, the anorthosite is complexly intruded by a light-pink muscovite-biotite granite.

Structure.—The anorthosite-norite body is bounded mostly by faults, but there are anorthosite dikes in the bordering gneiss. In the anorthosite, layering is not detectable; in the transition rock and norite, layering is prominent and dips outward.

Age.—The anorthosite-norite complex is younger than at least some of the adjacent gneiss. Neuerburg and Gottfried (1954) found, by the lead-alpha zircon method, ages of 930±90 million years for altered marginal norite and 810±80 million years for a pegmatite in the norite. Though these results are perhaps not wholly conclusive, the age of the complex may be taken tentatively as Precambrian.

Miscellaneous gneisses and associated rocks

Lithology and distribution.—Miscellaneous gneisses are mapped in every major division of the area east of the Newport-Inglewood line but are perhaps most abundant in the Transverse Ranges. These rocks vary greatly in lithology and may have a wide range in age.

The gneisses of the San Gabriel and San Bernardino Mountains are largely quartz plutonite gneisses, mostly of granitic composition in the western San Gabriels (Miller, 1934) and mostly quartz dioritic or granodioritic farther east (Alf, 1948; Hsu, 1955; Ehlig, 1958; Guillou, 1953; Allen, 1957). Locally there are bodies of pyroxene gneiss, hornblende gneiss, amphibolite, biotite quartz paragneiss, quartzite, and marble. A few lenses of granite and quartz diorite, up to 3 miles long, are mapped with the gneisses.

In the Peninsular Ranges, the gneiss belt near Palm Springs at the east edge of the San Jacinto Mountains is perhaps merely a marginal facies of the Mesozoic plutonic mass to the west (Fraser, 1931), especially as it has the same granodioritic to quartz dioritic composition. The other masses in the Peninsular Ranges, some 40 miles northeast of San Diego, are migmatitic bodies made up of quartz-rich sedimentary rocks of Bedford Canyon aspect (see below) and bands of the earliest pluton of the range, a granodiorite or quartz diorite (see Everhart, 1951; R. Merriam, 1958).

The gneiss areas of the Mojave Desert (Miller, 1946; Bowen, 1954) are mostly made up of quartz diorite, diorite, and granodiorite gneisses, more or less similar to those of the eastern San Gabriel Mountains, with some biotite schist, quartzite, and marble.

In general, the gneisses in most areas have quartz plutonite compositions, though less siliceous and more siliceous gneisses are locally voluminous and in the Transverse Ranges and Mojave Desert even carbonate rocks are present.

Structure.—In the San Gabriel Mountains the well-banded gneisses are typically made up of alternately light and dark bands ½ to 12 inches thick, the lighter bands mostly feldspar and quartz, the darker bands mostly feldspar, hornblende, and biotite. In many areas, as along San Gabriel Canyon below the

Forks or in the southeastern area studied by Alf (1948), the bands are less regular and up to several feet thick. The gneissic banding is variable in attitude, though Dibblee (unpublished), Ehlig (1958), and others have shown that approximately east-west strikes predominate near the south margin of the range and northwest strikes near the north margin.

In the western San Bernardino Mountains strikes are also predominantly east-west to northwest, fitting those in the nearest parts of the San Gabriel Mountains, as shown especially by Dibblee's unpublished mapping. In the southeastern San Bernardino Mountains the Mt. Pisgah ridge (lat. 34°+ and long. 117°—) also shows N. 70-75 W. strikes, though farther east Allen (1957) found varied attitudes, except in an area near long. 116° 30′, where north strikes and steep east dips prevail.

In the Peninsular Ranges, the strikes in the gneisses are mostly northwest. See especially Everhart (1951) and Merriam (1958).

Age.—In the western San Gabriel Mountains at least part of the gneiss is older than the anorthosite (Miller, 1934; Higgs, 1954) and so probably Precambrian. Elsewhere in the Transverse Ranges, and in the Mojave Desert, the ages of the gneisses are somewhat uncertain. The gneisses of the Peninsular Ranges may be wholly Mesozoic, derived in part from Cretaceous plutonites and in part from the Bedford Canyon formation, which contains Triassic fossils.

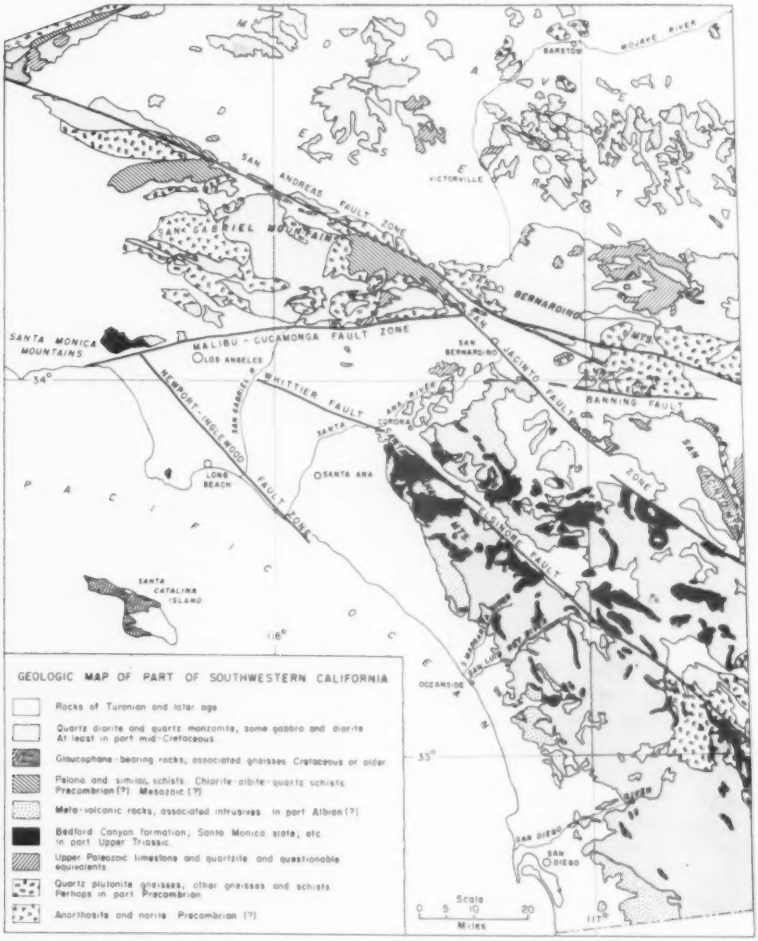
Known and questionable Upper Paleozoic rocks

North of Big Bear Lake in the northeastern San Bernardino Mountains the completely recrystallized Furnace limestone (Vaughan, 1922), which is mostly white, pure, and barren, contains a dark siliceous band with numerous fairly well preserved fossils. Collections made recently by Richmond (1954) have been assigned somewhat cautiously to the Mississippian by C. W. Merriam (1954, p. 13). Fusulinid ghosts collected at another locality in the limestone by Richmond are put tentatively in the Pennsylvanian (possibly Permian) by Merriam (1954, p. 13-14). On the Mojave Desert northeast of Victorville, limestone of the Oro Grande series has yielded a few determinable fossils of probable Pennsylvanian age and the clasts in an overlying, preplutonic limestone conglomerate have yielded numerous fossils of Permian and perhaps also Pennsylvanian age (Miller, 1944, p. 99-100; Bowen, 1954, p. 34 and 41; C. W. Merriam, 1954, p. 13). These fossils are the bases for the more or less confident assignment to the upper Paleozoic of crystalline limestones, quartzites, and other meta-sedimentary rocks with aggregate thicknesses of many thousands of feet, cropping out in the northern San Bernardino Mountains and in the Mojave Desert east and west of Victorville. McCulloh (1954, p. 17), however, has pointed out that the fossiliferous conglomerate overlying the Oro Grande series may be post-Paleozoic, perhaps Triassic.

Limestones, quartzites, mica schists, and gneisses (paragneisses?) in the eastern San Gabriel Mountains, eastern South Coastal Basin (south of San Bernardino), and in or near the San Jacinto Mountains are called Paleozoic(?), with great uncertainty. The evidence for the correlations, such as they are, is wholly lithologic. The best-studied mass (Baird, 1956; Ehlig, 1958) was divided

by Ehlig into 15 members with total thickness of about 7,300 feet. No correlation with the Furnace limestone or other formations of the northeastern San Bernardino Mountains seems possible.

Fig. 1.


SAN ANDREAS FAULT ZONE
SAN GABRIEL MOUNTAINS
SAN BERNARDINO MTS.
MALIBU - CUCAMONGA FAULT ZONE
SANTA MONICA MOUNTAINS
LOS ANGELES
NEWPORT - INGLEWOOD FAULT ZONE
WHITTIER FAULT
SAN JACINTO FAULT ZONE
BANNING FAULT
SAN JACINTO MTS.
SANTA ANA
LONG BEACH
SANTA CATALINA ISLAND
ELSINORE FAULT ZONE
SAN LUIS REY
OCEANSIDE
SAN DIEGO
VICTORVILLE
BARSTOW
MOJAVE RIVER
34°
33°
117°
118°
GEOLOGIC MAP OF PART OF SOUTHWESTERN CALIFORNIA
Rocks of Turonian and later age
Quartz diorite and quartz monzonite, some gabbro and diorite. At least in part mid-Cretaceous
Glaucophane-bearing rocks, associated gneisses. Cretaceous or older
Paloma and similar schists. Chlorite-albite-quartz schists. Precambrian (?). Mesozoic (?)
Meta-volcanic rocks, associated intrusives. In part Albian (?)
Bedford Canyon formation, Santa Monica slate, etc. In part Upper Triassic
Upper Paleozoic limestone and quartzite and questionable equivalents
Quartz plutonite gneisses, other gneisses and schists. Perhaps in part Precambrian
Anorthosite and norite. Precambrian (?)
Scale
0 5 10 20
Miles

Bedford Canyon formation, etc.

Lithology and distribution.—The Bedford Canyon formation of the Santa Ana Mountains, named by Larsen (1948), is mostly interbedded gray siltstones

and sandstones, which have been very slightly metamorphosed. The sequence dips east and was estimated by Larsen to be 20,000 feet thick. R. F. Yerkes and J. E. Schoellhamer tell me that they traversed some 18,000 feet of the formation in Silverado Canyon on the west side of the range and found about 59 percent siltstone, 40 percent sandstone, and 1 percent conglomerate and dark limestone. Some siltstones have original laminations 2 to 5 mm thick; others show poor slaty cleavage. The sandstones are graywackes, made up chiefly of quartz, rock fragments, and fine matrix, but with 2 or 3 percent feldspar.

Similar, gray to black rocks in the Santa Monica Mountains were called the Santa Monica slate by Hoots (1931, p. 88-89). The rocks were originally mostly sandy siltstone and silty sandstone; they are now somewhat metamorphosed and in a plutonic contact zone have been transformed into spotted slate.

East of the Santa Ana Mountains, in the northwest part of the Perris Block, a slightly metamorphosed, non-fossiliferous sandstone-siltstone sequence is assigned to the Bedford Canyon formation. A few miles farther east these rocks seem to become gneisses. Followed southeast along the strike, the sandstones and siltstones change gradually into schists (Hudson, 1922; Larsen, 1948). In the eastern part of the Perris Block, east of the gneiss belt mentioned above, recrystallized sandstones and siltstones of Bedford Canyon aspect are overlain by 20,000 feet of metamorphosed sedimentary rocks, now schist, quartzite, meta-arkose and meta-conglomerate (manuscript of H. P. Schwarcz, student at California Institute of Technology). All these rocks, and their probable equivalents southeast along the strike, are included with the Bedford Canyon formation in figure 1.

Structure.—In the Peninsular Ranges most of the rocks of this group strike northwest to north-northwest and dip northeast at moderate to steep angles. Little more than this is known about the structure of the type Bedford Canyon formation; even the top and bottom of the section are uncertain. Southeast of Perris (35-40 miles southeast of San Bernardino), Schwarcz found graded bedding and cross lamination that show most of this section to be an east-dipping homocline, right side up. In the highest part of the section, Schwarcz has followed key beds through folds. Durrell (1954) shows predominant east-west strikes in the Santa Monica Mountains and apparently a general anticlinal structure.

Age.—Triassic ammonites and other fossils have been found in the Bedford Canyon formation on both sides of the Santa Ana Mountains (Larsen, 1948, p. 18). At least one locality is Upper Triassic (letter of S. W. Muller to J. E. Schoellhamer). R. W. Webb (1939) reported a Mississippian coral found as float near the top of the metamorphic, possibly post-Bedford Canyon sequence southeast of Perris described by Schwarcz. Webb now thinks (letter of Nov. 20, 1959) that the coral may have been imported. Apparently influenced by Webb's fossil, Larsen (1948) mapped the metamorphic sequence as Paleozoic, even though it lacks the abundance of limestone, commonly in part very white, that is a characteristic feature of the Paleozoic rocks of the region.

In summary, the pre-Turonian rocks shown in black in figure 1 may all be Mesozoic or they may be partly Mesozoic and partly pre-Mesozoic.

Meta-volcanic rocks and associated intrusives

Lithology and distribution.—The Santiago Peak ridge, the highest part of the Santa Ana Mountains, is partially capped by gently dipping, gray volcanic breccias and flows, mostly andesitic, which are associated with fine to medium grained intrusives of similar or more siliceous composition. Larsen (1948) made this the type locality for the Santiago Peak volcanics, which extend in a discontinuous but broadening belt south-southeast to the Mexican border (fig. 1) and beyond. The largest intrusive body lies just east of the Santa Ana Mountains (Larsen, 1948, p. 36-39). Extrusives and intrusives are mapped together in figure 1.

In the northern Santa Ana Mountains the only quartz-bearing extrusive is a thin basal white tuff, but just south of the Mountains (along San Juan Creek) the west half of the belt is predominantly soda rhyolite tuff, soda rhyolite breccia, dacite, and the like. Farther south, near the Mexican border, siliceous volcanic breccias probably predominate, with siliceous tuffs, andesite, quartzite, slate, and metasedimentary schists also present. The degree of metamorphism varies with nearness to plutonic contacts and also geographically. Southeast of San Diego, the fine groundmass is almost everywhere completely recrystallized.

The bedrock volcanics of the Mojave Desert—the Ord Mountain group, and the Sidewinder and Hodge volcanic series (Gardner, 1940; Bowen, 1954)—are shown in figure 1 by the same pattern as the Santiago Peak volcanics because of similarities in lithology and stratigraphic position. The Sidewinder series also contains pink and red crystal-lithic indurated quartz-feldspar tuffs, locally with piedmontite, that have no analogues in the Santiago Peak volcanics but are very similar to clasts that are abundant in early Tertiary conglomerates unconformably overlying the Santiago Peak volcanics for almost the whole length of their California outcrop (Bellemin and Merriam, 1958; A. O. Woodford, ms.).

Structure and thickness.—The Santiago Peak volcanics are commonly obscurely or variably bedded. The original structures may have been complex and the rocks have surely been folded and faulted since deposition. Thicknesses of at least 2,000 to 3,000 feet are present in most areas (see, e.g., Hanna, 1926, p. 203) and greater thicknesses may occur near the Mexican border.

The Santiago Peak volcanics lie unconformably on the Bedford Canyon formation, in the Santa Ana Mountains with almost 90° difference in dip. They extend en masse downdip to the west beneath the superjacent series, as shown by well records, especially near Santa Ana and San Diego. They are not known as outliers east of the main outcrop belt. These facts indicate that the volcanics formed a sheet that has been eroded in the east, perhaps in part after deposition and some tilting of the Upper Cretaceous sediments, and has been protected from erosion in the west by downtilting and the presence of the superjacent series. The former eastern extension of the volcanics probably included reddish, piedmontite-bearing elements like those in the Tertiary conglomerates.

The Hodge series is stated by Bowen to be 10,000 feet thick and the Sidewinder series at least 4,000 feet thick. The Hodge series is homoclinal. The

Sidewinder rocks show complex interbedding of lenticular units, with variable, steep structures.

Age and correlation.—The Santiago Peak volcanics overlie the Bedford Canyon formation unconformably, as stated above. They are intruded by the plutonic rocks to be described below. In Baja California, 80 to 100 miles below the border, a series of andesitic breccias, tuffs, and flows, with some limestone, similarly intruded by plutonites, contains abundant gastropods, rare ammonites, and other marine invertebrates of Aptian and Albian ages (Allison, 1955; Larsen and others, 1958, p. 46; many unpublished data). Perhaps the Santiago Peak volcanics are also mid-Cretaceous.

The Ord Mountain, Hodge, and Sidewinder volcanic rocks are also intruded by plutonites. The Sidewinder series unconformably overlies a Permian or younger sedimentary formation.

Pelona and similar schists

Lithology and distribution.—In Sierra Pelona, a small Transverse Range north of the San Gabriel Mountains, the Pelona schist includes "nearly 7,500 feet of regularly bedded muscovite, chlorite-muscovite, and actinolite-chlorite schist," as well as "several hundred feet of quartzite, crystalline limestone, actinolite schist, chlorite schist, talc schist," etc. (Muehlberger and Hill, 1958, p. 634). The gray muscovite schists have a silky sheen, whose luster is in part due to a little graphite. Most of the schists contain nearly 50 percent plagioclase. The muscovite schists and the greenish-gray chlorite-muscovite schists contain about 15 percent quartz and may be meta-graywackes. The quartz-free actinolite-clinozoisite-chlorite-plagioclase schists may be meta-spilites.

The second large mass of Pelona schist, in the eastern San Gabriel Mountains, is about 10,000 feet in maximum thickness (Ehlig, 1958). It is made up of 3 rock types—gray schists with abundant quartz and muscovite, which may be meta-graywackes or meta-tuffs; green schists, composed of albite, epidote, chlorite, and green or blue-green amphibole, which may be meta-spilites; and laminated quartzites that are considered meta-cherts. Ehlig has divided the formation into 3 members, characterized by different proportions of the 3 rock types.

Strips of Pelona-like schist are also present northeast of the San Andreas fault in two places, one just north of the Sierra Pelona mass, the other in the Tehachapi Mountains at the northwest corner of figure 1. The Tehachapi mass is mostly mica-chlorite-quartz-albite schist, amphibole-albite schist, and quartzite, 5,000 feet thick, nearly vertical, and striking northeast (Wiese, 1950, p.12).

The Pelona schist is probably not present south of the Malibu-Cucamonga fault zone west of the San Jacinto fault. The suggestion of its presence in the bedrock beneath the Puente Hills, just northwest of the Santa Ana Mountains (Schoellhamer and Woodford, 1951), has not been borne out by cores from several widely spaced later wells. These cores indicate that mildly metamorphosed, andesitic(?) volcanic rocks, possibly the Santiago Peak volcanics, are widespread beneath the Puente Hills just northeast of the Whittier fault.

Structure.—In Sierra Pelona the schist is folded, with the principal structure an anticline that trends about N. 70° E. In the eastern San Gabriel Moun-

tains the structure is complex, in part homoclinal with southwest dips; the section is topped by mylonite and gneiss. Nowhere is the base of the schist exposed.

Age and correlation.—The age is unknown. A Precambrian age has been suggested by Miller (1946), among others. Ehlig (1958) favors a late Mesozoic post-batholithic age, because he considers the fine grained quartz monzonite intrusions in the schist atypical. Correlation with the Rand schist north of the Mojave Desert area shown in figure 1 has been suggested by Hulin (1925) and others. Correlation with the Orocopia schist of southeastern California was suggested by Hill and Dibblee (1953, p. 450).

Glaucophane-bearing rocks, etc.

Lithology and distribution.—A metamorphic facies of the Franciscan formation, on Santa Catalina Island, and the Catalina schist of the mainland, west of the Newport-Inglewood zone, both contain glaucophane and lawsonite, minerals not found elsewhere in the area of figure 1, though amphiboles related to glaucophane occur in the eastern San Gabriel Mountains, sparingly in the Pelona schist (Ehlig, 1958) and very rarely with Paleozoic(?) lime silicates (Laudermilk and Woodford, 1940, p. 418). On Santa Catalina Island, there are also higher rank metamorphic rocks without glaucophane or lawsonite.

The Franciscan formation of Santa Catalina Island includes metasandstone with 1 to 30 percent lawsonite and glaucophane, fine grained glaucophane schist derived from shale, quartzite probably derived from chert, rare green or red chert with clay partings of typical Franciscan type, metaconglomerate, meta-volcanic rocks (andesites and spilites?) with glaucophane, talc-actinolite schist in sill and dike forms, and serpentine bodies (Bailey, 1941; Woodford, 1924). The Franciscan rocks are overlain by hornblende gneiss, quartz muscovite gneiss, and a chlorite-talc-actinolite melange.

The Franciscan(?) Catalina schist of the mainland crops out in the Palos Verdes Hills west of Long Beach (Woodring and others, 1946) and is the only basement rock encountered in wells in the Los Angeles Basin west of the Newport-Inglewood zone.

> The Catalina schist in the Los Angeles Basin is composed chiefly of fine-grained gray-green chlorite-bearing schists, such as muscovite-chlorite schist, muscovite-chlorite-quartz schist, and albite-chlorite-quartz schist . . . Glaucophane and lawsonite . . . are widespread but less abundant. (Schoellhamer and Woodford, 1951.)

Just north of Long Beach saussurite gabbro, a natural associate of glaucophane schists, was found within the Newport-Inglewood fault zone. Southeastern extension of the schist for at least 65 miles is suggested by a thick Miocene breccia (Woodford, 1925) composed of Catalina schist blocks. Amphibole-epidote-albite schists (probably derived from basic volcanic rocks) and saussurite gabbro are locally abundant in the breccia.

No quartz-bearing plutonic rock has been found in or adjacent to the Franciscan rocks or the Franciscan(?) schists. A fine-grained intrusive that makes up much of the eastern part of Santa Catalina Island was called a hornblende quartz diorite porphyry by Bailey (1941), but it has chilled margins, "has had no appreciable effect on the earlier metamorphic series" (Bailey, 1941, p. 18-

19), and is better called a dacite porphyry. The representation of this rock by Jahns (1954b) in the same pattern as the rocks of the southern California batholith is misleading. For one thing, the Santa Catalina Island rock may be considerably younger.

Structure.—On Catalina, the Franciscan rocks show many undistorted relic sedimentary structures, among which the shale flakes in the meta-sandstones are notable. The general structure is somewhat complex, but a broadly synclinal structure has been made out, with high-grade hornblende and other schists and gneisses on top and in the center of the syncline, perhaps as a result of thrusting.

On the mainland, attitudes range from horizontal to vertical, with dips less than 20° most common. The detailed structure is crumpled and lineated, and the general structure is probably complex.

Age and correlation.—My impression that the non-schistose meta-sedimentary rocks on Santa Catalina Island are Franciscan has strengthened with the years, especially as a result of Bailey's (1941) work and the discovery that Franciscan is a facies rather than an age term (Irwin, 1957). I feel much less confident about the mainland schists, especially now that more typical Franciscan has been found far to the south in the San Benito Islands and in Cedros Island off central Baja California (van West, 1958). The mainland schists are thoroughly foliated, lack Franciscan relics, and are associated with saussurite gabbro instead of serpentine. They may belong to the same sequence as the Catalina Franciscan and so be about the same age, or they may not.

Quartz plutonites, etc.

Lithology and distribution.—Quartz plutonites are important to dominant elements of the bedrock complex everywhere east and north of the Newport-Inglewood zone.

The plutonic rocks of the Peninsular Ranges have been described in considerable detail by Larsen and his students (Larsen, 1948; Hurlbut, 1935; F. S. Miller, 1937; Everhart, 1951). These and previous workers have named at least 6 different tonalites (quartz diorites), some of which occur in many separate outcrop areas, 6 granodiorites, 2 granites, and 1 gabbro (extensive). In the San Jacinto Mountains Fraser (1931, p. 526) found quartz diorite and granodiorite perhaps twice as abundant as granite and quartz monzonite.

In the western San Gabriel Mountains Miller (1934) found a dark quartz diorite and a pale granodiorite, both common. Dibblee (unpublished map) extended the pale rock (typically 2 to 4 percent dark minerals, rarely up to 10 percent) east and northeast over a large area and found the average rock quartz monzonite. A somewhat similar quartz monzonite is the prevalent rock in the San Bernardino Mountains and Mojave Desert.

The plutonic rocks of the Peninsular Ranges extend south into northern Baja California, where a sample area in which quartz diorite and granodiorite predominate has been described by Woodford and Harriss (1938).

Structure.—In the Peninsular Ranges the individual plutons have many shapes, but most are more or less equidimensional, rather than greatly elon-

gated. A few are arranged in concentric bands (Larsen, 1948, pl. 1; R. Merriam, 1941). A widely distributed quartz diorite and some other plutons show banding and orientation of gneissic and gabbroic inclusions, with prevalent northwesterly strikes, roughly parallel to the structures in the enveloping metamorphic rocks (see R. Merriam, 1958, fig. 2).

In the eastern San Gabriel Mountains, relatively small bodies of quartz diorite are greatly elongated sill-like masses with internal structures roughly parallel to their boundaries and to the structures in the surrounding gneisses. In the southeasternmost part of these mountains Alf (1948) found an east-west trending belt of dark gray quartz diorite mylonite gneiss 13 miles long and some 2,000 feet thick, bounded on the north by dark gneissoid quartz diorite. These structures are in striking contrast to those of the Peninsular Ranges.

The plutonic structures of the San Bernardino Mountains and the Mojave Desert are only locally well known.

Age and correlation.—Quartz plutonites intrude all other pre-Turonian rocks in all the areas of their occurrence. The intrusions having these relationships are therefore the youngest part of the bedrock. A few plutonic masses may be older, perhaps much older.

The southern California batholith of Larsen (1948) extends throughout the Peninsular Ranges of southern California and northern Baja California. About 120 miles south of the border, the San Jose quartz diorite (Woodford and Harriss, 1938, p. 1314) intrudes a volcanic and sedimentary series that contains *Douvilleiceras* very close to *D. mammillatum* (Larsen and others, 1958, p. 46), a middle Albian guide. Here the plutonite is certainly post-middle Albian and, as indicated in the introduction, probably pre-Turonian. Larsen and others (1958, p. 47) give 4 lead-alpha ages for the San Jose and neighboring plutons, with a range of 98-112 million years and a mean of 103 million years.

Larsen and others (1958, p. 47-50) give about 20 lead-alpha ages for quartz plutonites from various parts of the Peninsular Ranges in southern California, with a range from 92 to 136 million years and a mean near 109 million years. The same authors give one determination for the San Gabriel Mountains, for a quartz diorite from the top of Mt. Wilson (1958, p. 48), with an age of 122 million years, well within the range for Peninsular specimens. Several determinations for plutonic rocks in the eastern Mojave Desert and east of the San Bernardino Mountains are also in the same range.

Larsen and others (1958, p. 52-53) also give 15 lead-alpha ages for plutonites collected farther north in California, mostly in the Sierra Nevada, ranging from 81 to 116 million years, with a mean of 102 million years. By this method, the Peninsular, Transverse Range, Mojave Desert, and Sierra Nevada plutonites all seem to have about the same age. Other geochemical data (Jahns, 1954a, p. 25; Curtis and others, 1958) are roughly similar; some variations are important but apparently not significant in the present connection.

MAJOR FAULTS OF THE SOUTH COASTAL BASIN

In the South Coastal Basin, some major faults cut both bedrock and superjacent series. All are steep faults, with dips of 65° or more. The Basin is

bounded on the north by the Malibu-Cucamonga fault zone and on the northeast by the San Andreas fault zone. Each zone includes some faults bounded on both sides by sediments. The San Jacinto, Newport-Inglewood, and Whittier faults cut across the Basin (fig. 1). The Malibu-Cucamonga zone cuts off the Newport-Inglewood zone and so may be the younger. The San Jacinto fault cuts off the Malibu-Cucamonga zone and so might be considered still younger. But movements on all three zones have extended into recent geologic time, with epicenters of major historic earthquakes on the Newport-Inglewood as well as on the San Jacinto faults.

In this basin, as elsewhere in the region, faults show geomorphic evidence of late dip-slip and strike-slip movements. The late Quaternary displacements near the east end of the Malibu-Cucamonga zone may be entirely dip slip (Eckis, 1928), reverse faulting. Quaternary strike slip on the Whittier fault may amount to thousands of feet, as suggested by offset streams.

The many stratigraphic units in the superjacent series make possible detailed histories of the structural evolution—folding, faulting, development of unconformities—along some faults. But the abundance of information also makes an acceptable interpretation very difficult. Detailed mapping along the Whittier fault by the U. S. Geological Survey, 1945-59, even now has not completely clarified the history of the displacements. It does seem possible to say that the maxima for dip slip and strike slip are almost certainly each less than 15,000 feet in the superjacent series. Some fault movement may have occurred in the bedrock before the Turonian deposition, with Santiago Peak volcanics dropped a few thousand feet on the southwest side and brought against quartz diorite, porphyry, and Bedford Canyon formation to the northeast. Strike slip might possibly give the same results, though the Santiago Peak volcanics(?) beneath the Puente Hills count against such an interpretation.

DISCUSSION

Three bedrock belts

One type of bedrock can be isolated immediately. It lies west of the Newport-Inglewood fault zone and south of the Malibu-Cucamonga fault zone. It is characterized by the presence of glaucophane-bearing rocks and the absence of quartz plutonites and quartz plutonite gneisses.

The remainder of the map area can be divided less sharply into two parts, one with graywackes of Bedford Canyon type and almost without limestone, the other with late Paleozoic or late Paleozoic(?) limestones but no rocks of the Bedford Canyon association. The black patches on figure 1 show that Bedford Canyon and related rocks are present in the Santa Monica Mountains of the Transverse Ranges and in all parts of the Peninsular Ranges except the San Jacinto Mountains. Late Paleozoic limestones are present in the Mojave Desert and San Bernardino Mountains. Similar but non-fossilliferous limestones are present in the San Gabriel Mountains, San Jacinto Mountains, and the east end of the South Coastal Basin. The Pelona schist occurs exclusively in this third belt but is not closely associated with the limestones.

All three belts may belong to a eugeosynclinal zone, though the first quali-

fication for such zones given by Stille (1940, p. 653), namely, that they are "the highly mobile zones of the geosynclinal systems, in general the first to be folded", cannot be ascertained. All the belts contain magmatic elements and so possess Stille's second qualification. Nevertheless, the limestone-quartzite formations represented in figure 1 appear to be lacking in igneous elements and the great thicknesses of quartz-rich Bedford Canyon sediments nearly so. These two sets of stratigraphic units may therefore be considered useful guides to atypical belts in this particular eugeosynclinal zone.

Variations in metamorphism

The meta-sedimentary and meta-volcanic rocks show great variations in metamorphic effects. The type Bedford Canyon sandstone is unfoliated and only its matrix is recrystallized. Most other rocks of the middle belt also show relatively little foliation, even where completely crystalloblastic. A traverse beginning at the northwest corner of the Perris block, just east of the Santa Ana Mountains, and extending southeast for 35 or 40 miles would show changes in mineralogy suggestive of increasing temperature and pressure. A narrow chlorite zone might be succeeded by a broad biotite zone and finally by rocks with sillimanite. But the large area with well preserved sedimentary textures and structures, studied by Schwarcz (ms.), is near the southeast end of this traverse, where the minerals indicate the greatest metamorphism.

The Franciscan sandstone of Catalina Island and the fossiliferous Furnace limestone of the San Bernardino Mountains are both wholly crystalloblastic, but typically neither is foliated or distorted. The sandstone has preserved the forms of included shale flakes and the limestone the shapes of invertebrate fossils.

The Catalina schist of the Los Angeles Basin and the Pelona schist of the northeastern belt are foliated and crystalloblastic. The Pelona schist of the San Gabriel Mountains locally preserves some relic sedimentary structures.

The Santiago Peak and Sidewinder volcanics are mostly only partially recrystallized. They are only locally foliated.

Correlations

The statements concerning age and correlation in the previous sections show how little is known about the ages or even the local correlations of the sedimentary and volcanic rocks. The batholithic rocks, on the other hand, are extensively and usefully dated, though in the area of figure 1 the results are not precise.

Correlations between belts, or even between different parts of a single belt, are precarious. We might start with the Santa Catalina Island Franciscan, take its age tentatively as Jurassic or Cretaceous or both, by analogy with the type Franciscan (Irwin, 1957), and then correlate even more tentatively with the Catalina schist of the mainland and even the Pelona schist of the northeastern belt, skipping the middle belt altogether. These correlations would make the schists younger than the intervening, slightly slaty Bedford Canyon formation. Determined correlators might rely on analogy with the increase of metamor-

phism upsection in the Perris Block (Schwarcz, ms.), but probably would not rely on plutonic intrusives to furnish heat for the increase in metamorphism. Or strike slip faulting might be called upon, or even nappe thrusting. Another tactic would be to give up the attempt to correlate. At the least, one may say that in this region correlation from belt to belt is difficult, except for the late igneous rocks, both volcanic and plutonic, in the central and northeastern belts.

Extent and shape of the California-Baja California Cretaceous batholith

The geologic map of North America (Stose, 1946) or Larsen's southwestern plutonic map (1954, fig. 1), if supplemented by Dibblee's data (included in fig. 1 of this paper), would show a nearly continuous strip of Cretaceous quartz plutonites from the Sierra Nevada of northern and central California to the Sierra San Pedro Mártir of north central Baja California. The strip is many hundreds of miles long and 50-75 miles wide; it strikes about N. 30° W. In and northwest of the Transverse Ranges the plutonic outcrops widen greatly and the concealed portions of the batholith may be even more extensive. However, the N. 30° W. strip may reasonably be considered the principal body, or core, with a deep root extending from the Sierra Nevada into the Peninsula of Baja California and making possible the mountainous surface expression.

Problems of the Transverse Ranges

The principal structural anomaly in southern California is represented by the Transverse Ranges, whose elongation and structures trend approximately east-west. Not only do the structures north and south of these ranges trend northwest, but the main rock belts of the region extend more or less northwest-southeast, even cutting across the Transverse Ranges in a west-northwest direction, as shown in figure 1. Gneisses and thick mid-Cretaceous (and older?) mylonite zones seem to characterize the Transverse Ranges. In the central and eastern parts of the South Coastal Basin some structures suggest transitions between Peninsular and Transverse trends. Similar transitions occur north of the Transverse Ranges. These ranges form a kink that complicates all the structural problems of the region.

Strike slip on the San Andreas fault

In the San Francisco earthquake of 1906 right lateral strike slip on the San Andreas fault was as much as 22 feet in central California. In the area of figure 1, offset streams indicate that right lateral movement amounting to thousands of feet occurred during the Quaternary. For pre-Quaternary movements the distinction between separation and slip must be made (M. L. Hill, 1959; Crowell, 1959). Noble (1954) found a structurally complex succession of granodiorite, Paleocene, and Miocene north of the San Gabriel Mountains offset in a way that seems to require 30 miles of right lateral strike slip since the Middle Tertiary.

Greater displacements may have occurred in the bedrock but they have not been demonstrated. Some apparent net separations are puzzling, to say the least. In figure 1 the arrangement of the 4 principal masses of Pelona and similar schists seems to show *left* lateral separations of 40 miles along the San Andreas fault. The present pattern might be explained by Mesozoic left lateral

movement of 70 miles, followed by Cenozoic right lateral movement of 30 miles. However, comparison of the 2 northwestern Pelona masses, one in Sierra Pelona (Muehlberger and H. S. Hill, 1958) and the other in the Tehachapi (Wiese, 1950) brings out structural differences sufficient to cast doubt on the matching of the two masses.

M. L. Hill and Dibblee (1953, p. 450) cautiously suggested offset of the San Gabriel Mountains Pelona mass from the Orocopia schist on the opposite side of the fault, 160 miles southeast (outside figure 1). This separation would be right lateral. A slip of 160 miles would mean a displacement of the Sierra Nevada-Peninsular core of the Cretaceous batholith from an anomalous disjunction to the present alignment. It would not decrease the zigzag in the quartz diorite line shown by Moore (1959, p. 199) but would on the contrary make the zag more extreme. It would mean that the Santiago Peak volcanics at time of deposition were completely out of line with their apparent correlatives on the Mojave Desert, the Sidewinder volcanics. It would move the Santa Catalina Island-San Benitos Island-Cedros Island Franciscan (van West, 1958) to an original position east of the main mass of Franciscan rocks in central and northern California. The east edge of the southern California and Baja California Franciscan and related rocks is already a little too far east to fit the general Franciscan boundary farther north (Reed, 1933, fig. 6, p. 28), possibly as a result of left lateral strike slip on the Malibu-Cucamonga fault zone. Possibly the present position of this boundary (the Newport-Inglewood line) is the resultant of a large right lateral movement on one fault and a large left lateral movement on another. But the violence done to the Franciscan boundary is no recommendation for a 160-mile strike slip on the San Andreas zone.

Muehlberger and H. S. Hill (1958, p. 640) have pointed out that "the present wide distribution [of the Pelona schist] may be the result of original deposition over a large area rather than dislocation by later tectonic events (for example, by major strike-slip faulting)". But the termination of 3 of the 4 principal Pelona masses at the San Andreas fault is a feature that demands explanation. This difficulty will be discussed below.

Suggestions of strike slip much greater than 160 miles, on the San Andreas and other faults, have been made by Curtis and others (1958, p. 16) and by King (1959, fig. 95). These authors propose that one or both of the granitic strips that extend, respectively, 75 miles west and 300 miles northwest into the California coastal Franciscan terrane (cf. Reed, 1933, fig. 6, p. 28) originated as parts of the Cretaceous plutonic strip mentioned above and were then transported bodily into their present positions along several faults. King's highly generalized diagrams (1959, figs. 92 and 95) seem to show that one may preserve the alignment of the batholithic core and have great strike slip too, if segments of the batholith are sliced properly and moved right and left, as desired. No detailed specifications for the movements have yet been presented. When the specifications are available they may be tested roughly by the Olaf P. Jenkins edition of the California geologic map, now being issued in parts by the California Division of Mines. The specifications should also be a stimulus to more detailed mapping that will make a critical evaluation possible. At best,

evaluation may be difficult. In fact, King considers the sequence of geologic events so complicated that, "like Humpty-Dumpty, all the King's horses and all the King's men cannot put the Coast Ranges back together again" (1959, p. 171). Possibly other horses and other men will have no better luck.

Under the circumstances, I prefer a working hypothesis that includes some dip slips and limits the strike slip on the San Andreas to 30 miles, right lateral. The northeastern continuations of the Sierra Pelona and San Gabriel Mountains schist masses would be raised high and left in the air over the southern Mojave Desert and San Bernardino Mountains, except for the little patch reported by Allen (1957). The third Pelona schist mass apparently truncated by the San Andreas fault is so long and narrow that it may be only a local roof pendant in the Cretaceous batholith (or, if post-batholithic, a narrow sliver beneath an ancient thrust exposed not far away). The gneiss and plutonic masses in the San Gabriel and San Bernardino Mountains seem to fit as well as could be expected considering the postulated dip slip, though final judgment must be reserved until the southern San Bernardino Mountains are more thoroughly mapped geologically.

The tentative choice of short slips, if these will do the business, is an example of the use of principle *Disjunctiones minimae, disjunctiones optimae*. This rule may be considered a quantitatively parsimonious relative of Ockham's law: *Entia non sunt multiplicanda praeter necessitatem*.

Note that the questions raised here have concerned the magnitude of strike slip, not its existence, and the difficulty of determining, in a given case, whether displacement has been strike slip, dip slip, or oblique slip. Emphasis has also been placed on the importance of rock correlation, in the stratigraphic sense, before structural interpretation is attempted. Finally, full understanding can only follow complete historical analysis. One might discover, for example, the following sequence of structural events: folding, production of an unconformity, vulcanism, deformation and mild metamorphism, plutonism and extreme local metamorphism, production of major unconformity, very complex local faulting, and finally moderate dip slips and greater strike slips on steep master faults. But the sequences in different subdivisions of this region cannot be compared with any confidence unless the events in each subdivision can be at least roughly dated in terms of a general time scale. If one goes beyond a specific problem, such as strike slip on the San Andreas fault, and attempts to reconstruct the geologic history of the whole region, dating becomes the most serious problem for bedrock studies in southwestern California.

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PALEOECOLOGIC DISSONANCE; *ASTARTE* AND *NIPA* IN THE EARLY EOCENE LONDON CLAY*

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ABSTRACT. *Astarte* (a chiefly circumarctic and boreal marine pelecypod) and *Nipa* (an Indo-Malayan palm) lived in the same region during early Eocene time when the London clay was deposited, whereas now they are separated by a gap of several thousand kilometers. Such examples of paleoecologic dissonance are more common than indicated by paleoecologic generalities concerning Teriary marine faunas.

The London clay, of early Eocene age, has yielded a remarkable flora of pyritized fruits and seeds, described in the superb monograph by Reid and Chandler (1933). Their material was gathered on the beach of the Isle of Sheppey, on the south side of the London basin near the mouth of the Thames. Though the beach is backed by cliffs formed by the London clay, Mrs. Reid and Miss Chandler had no outcrop specimens. Since then a few fruits and seeds have been found in place in the cliff (Davis, 1936, p. 341-342). They are, however, very rare as compared with their abundance in the beach concentrates.

The affinities of the flora are tropical, overwhelmingly Indo-Malayan (Reid and Chandler, 1933, p. 47-62). Fruits, or fragments of fruits, representing the same genus as the Indo-Malayan nipa palm (*Nipa*) are the most abundant of these fossils. No fruits of mangrove, with which the modern nipa palm is associated, were available to Reid and Chandler, but that deficiency was later remedied by Miss Chandler (1951). Reid and Chandler concluded that the London clay plants grew on the shore of an arm of the Tethys Sea. As a matter of fact, remains of *Nipa* have been found in Eocene deposits at other localities along the shores of Tethys and its arms: in Belgium, France, Italy, southern Russia, Senegal, Egypt, and Borneo (Reid and Chandler, 1933, p. 130).

The London clay also contains a large fauna of marine invertebrates, including some 150 genera of mollusks. Some of the mollusks have Tethyan affinities, but more have northern affinities (Davis and Elliott, 1958, p. 265). The anomid genus *Enigmonia* (Wrigley, 1936), which now lives on roots, branchlets, and leaves of mangrove and nipa palm, is the most striking of the Tethyan genera.

Astarte may be taken as an example of the northern genera. Several generations of competent paleontologists have studied the fauna of the London clay. They agree that, so far as shell features are concerned, the fossils are indeed to be referred to the modern genus *Astarte*. Astarte may not have been a frigid lady—in fact, she was anything but frigid, if the stories about her can be believed—but the clam that bears her name is at the present time chiefly circumarctic and boreal. In the Pacific Ocean *Astarte* ranges southward to Puget Sound and northern Japan. In Atlantic waters, however, it extends to the Mediterranean Sea and the Caribbean Sea, but, as usual with northern genera extending far southward, the southern species are found in fairly deep water. Like many Arctic genera, *Astarte* also occurs in Antarctic waters.

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According to Davis and Elliott (1958, p. 270), an early transgression of the London clay sea from the northeast brought in the northern forms and later a Tethyan connection from the southwest was established. Even during the early transgression, however, tropical plants, including *Nipa*, were growing along the coast. After the Tethyan connection was established, the part of the London clay exposed in the Sheppey cliffs was deposited, and both *Astarte* and *Nipa* are found there (Davis, 1936, p. 337, 341-342).

Astarte and *Nipa* lived in the same region during early Eocene time, but now are separated by a gap of several thousand kilometers. One or the other has evolved in adaptive qualities. Inasmuch as the London clay flora is overwhelmingly tropical, *Astarte* evidently is the one that has evolved.

The Eocene occurrence of *Astarte* and *Nipa* is an example of paleoecologic dissonance—an example to which Wrigley (1949, p. 231-232) pointed some years ago. Such examples are more common than indicated by paleoecologic generalities concerning Tertiary marine faunas.

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HIGH-PRESSURE FORM OF ANALCITE AND FREE ENERGY CHANGE WITH PRESSURE OF ANALCITE REACTIONS

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ABSTRACT. The compressibility of four natural analcites has been obtained up to 10,000 atm at 25°C. All specimens show abnormal compressibility $[(\partial\beta/\partial P)_T > 0]$ and one exhibits a reversible transition at about 8400 atm. Abnormal increase of compressibility with pressure usually precedes a transition, and it is suggested that a similar transition lies at higher pressures than those investigated for the three remaining specimens.

The compressibility and density data were used with previously obtained data to compute the volume change and free energy change with pressure for the reactions 2 analcite $\rightleftarrows$ nepheline + albite + 2 water, analcite $\rightleftarrows$ jadeite + water, and analcite + quartz $\rightleftarrows$ albite + water. The volume change for these reactions proceeding to the right is negative for all pressures investigated at 25°C.

The analcite exhibiting the transition showed an increase in birefringence after compression, which is attributed either to strain induced by the volume discontinuity or to retention of a possible lower symmetry of the high-pressure form. Lower symmetry in a high-pressure form is not uncommon in previously investigated substances.

INTRODUCTION

The stability ranges of minerals have been calculated from their thermodynamic properties in order to obtain a guide for experiment or as a test of experimental results. One mineral, jadeite, has been the subject of intense study for many years because of its bearing on problems relating to minerals and rocks believed to have formed at high pressures. Prior to the successful synthesis (Coes, 1955) and determination of portions of the field of stability of jadeite (Griggs and Kennedy, 1956; Robertson, Birch, and MacDonald, 1957), thermochemical data were collected for the purpose of computing the stability field of jadeite. The principal reactions for which such data were obtained are

$$\text{nepheline} + \text{albite} = 2\ \text{jadeite} \qquad (1)$$

$$\text{albite} = \text{jadeite} + \text{quartz} \qquad (2)$$

The large amount of data collected by various workers for this purpose has been summarized by Adams (1953) and by Fyfe and Valpy (1959). In addition to these anhydrous reactions, others involving water also limit the field of stability of jadeite:

$$2\ \text{analcite} = \text{nepheline} + \text{albite} + 2\ \text{water} \qquad (3)$$

$$\text{analcite} = \text{jadeite} + \text{water} \qquad (4)$$

$$\text{analcite} + \text{quartz} = \text{albite} + \text{water} \qquad (5)$$

Efforts were made, therefore, to acquire the additional data, namely the thermochemical properties of analcite, necessary to compute these reactions. The low-temperature heat capacity and entropy at 25°C of analcite were obtained through the courtesy of Dr. K. K. Kelley by King (1955) and preliminary data on the heat of solution were obtained through the kindness of Dr. F. C. Kracek (personal communication, 1954). The compressibility, high-temperature specific heat, thermal expansion, and specific volume data are still to be

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measured. In the meantime, reaction (3) has been investigated experimentally (Yoder, 1950, 1954; Valpy, 1958) at relatively low pressures; reaction (4) was studied at relatively high pressures (Griggs and Kennedy, 1956; Robertson, Birch, and MacDonald, 1957); and data have been obtained by Ellis (1958) and Valpy (1958) on reaction (5).

It is the purpose of the present paper to record the compressibility and density data for several natural analcites. These data are required for calculating the free energy change in reactions (3), (4), and (5) as a function of pressure.

APPARATUS AND SPECIMENS

The apparatus for compressibility measurements has been previously described by Weir (1950), and the equations for calculation have been outlined by Adams, Williamson, and Johnston (1919). The specimen is immersed in Varsol in a pressure vessel fitted with a leak-proof piston. The displacement of the piston is measured at a series of pressures up to 10,000 atm. A similar experiment performed with a steel bar of known compressibility permits the calculation of the compressibility of the specimen.

A few additional comments are appropriate with regard to the calibration, by means of the Ice VI$\rightleftarrows$L transition, of the manganin-resistance pressure gage used in the apparatus. The temperature of the room in which the equipment was installed was regulated to 20° ± 0.5°C, and the temperature variation in the apparatus is presumed to be less than 0.1°C, so the value used for the Ice VI—water transition was 8710 atm (Bridgman, 1912). For thermodynamical purposes the accuracy of this value is not very great (Bridgman, 1912; Adams, 1931), since the value is extremely sensitive to changes in temperature (0.1°C = 15.4 atm). An appreciation of the precision may be gained from figure 1. The values given by Bridgman have been generally accepted subject to later correction, if any. For the present study, errors as great as ± 20 atm have negligible influence on the accuracy of the volume changes measured.

The specimens of analcite investigated were selected on the basis of composition or availability of especially large single crystals. Two specimens (Golden, Colo., and Wasson's Bluff, Nova Scotia) consisted of crystal fragments and were contained in an open, thin steel vial. (The compressibility of steel has been determined by Bridgman, 1940, 1949.) Two other specimens (Seisser Alp, Austria, and Fassa Thal, Austria) were used in the form of cores drilled from exceptionally large single crystals. The core assembled from each specimen was 1.5 cm in diameter and approximately 10 cm long. No container was required for these specimens.

Representative chemical analyses of specimens from the same localities are given in table 1. These analyses are 1.47 to 2.48 percent higher in SiO_2 than the ideal composition, a common feature of analyzed analcites (see Saha, 1959, p. 310, fig. 4). These high SiO_2 values cannot be attributed to improper analytical technique because the $Al_2O_3 + SiO_2$ totals are also higher than expected. The material from Golden, Colo., purified by the writers, was analyzed with extreme care by E. H. Oslund (Rock Analysis Laboratory, Univer-

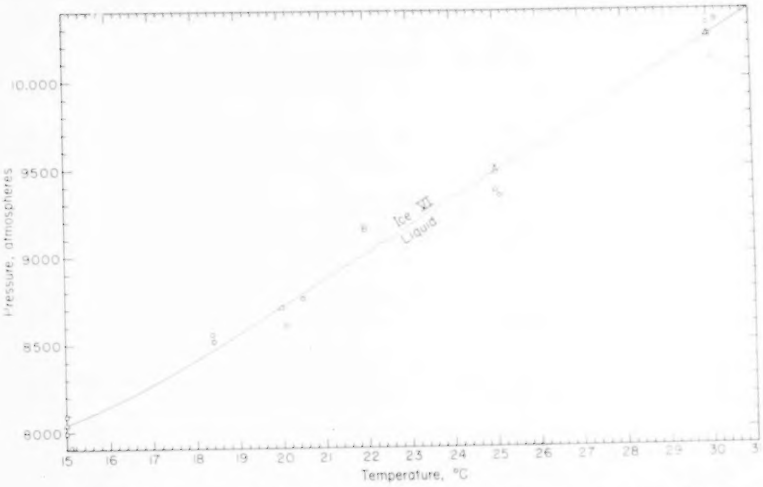

Fig. 1. Melting curve for Ice VI used as a standard for pressure gage calibration. Circles are data of Bridgman (1912) and triangles his preferred values at fixed points; cross is data of Adams (1931, p. 3778). The extreme sensitivity of the melting behavior to temperature and the dispersion of the data are emphasized.

TABLE 1

Representative chemical analyses of analcites from the same four localities as those investigated

	Ideal	A Golden Colorado (King, 1955)	B Wasson's Bluff Nova Scotia (Clarke and Steiger, 1899)	C Seisser Alp Austria (Baschieri, 1908)	D Fassa Thal Austria (Lemberg, 1876)
SiO_2	54.58	56.05	57.06	56.49	56.32
Al_2O_3	23.16	22.36	21.48	21.89	22.00
Fe_2O_3		0.03	0.13	—	—
FeO					
MgO		0.02	—	—	—
CaO		0.00	0.16	0.56	0.51
Na_2O	14.08	13.44	12.20	11.39	13.19
K_2O		0.10	—	0.53	—
H_2O^+	8.18	8.13	8.38	8.73	8.80
H_2O^-		0.01	0.58		
TiO_2		0.01	—	—	—
Total	100.00	100.15	99.99	99.59	100.82

A. $(Na_{0.93}Mg_{0.01})(Al_{0.99}Si_{2.01})O_6 \cdot 0.98\ H_2O$
B. $(Na_{0.89}Ca_{0.01})(Al_{0.92}Si_{2.08})O_6 \cdot 1.02\ H_2O$
C. $(Na_{0.80}Ca_{0.02}K_{0.03})(Al_{0.96}Si_{2.05})O_6 \cdot 1.06\ H_2O$
D. $(Na_{0.92}Ca_{0.03})(Al_{0.93}Si_{2.06})O_6 \cdot 1.06\ H_2O$

sity of Minnesota) for use in correcting the thermochemical data subsequently obtained on portions of the same sample by King and by Kracek. No excess SiO_2 as quartz was observed in the sample either optically or by means of x-ray diffraction patterns. (From another locality one analyzed analcite known to have a high SiO_2 content was shown as a result of later x-ray study to contain quartz.)

The analcites having high silica contents are described as belonging to an extensive solid solution series ranging from $Na_2O \cdot Al_2O_3 \cdot 2SiO_2 \cdot H_2O$ hydrous "nepheline" composition) to $Na_2O \cdot Al_2O_3 \cdot 6SiO_2 \cdot 3H_2O$ (hydrous "albite" composition). Assuming that the Al and Si ions are distributed over the same position of the analcite structure as assigned by Taylor (1930, 1938), the following general formulas may be deduced for the molecular ratios of interest:

1:1:2:1	$Na_{1.5}(Al_{1.5}Si_{1.5})O_6 \cdot 0.75\ H_2O$
1:1:3:1.5	$Na_{1.2}(Al_{1.2}Si_{1.8})O_6 \cdot 0.90\ H_2O$
1:1:4:2	$Na_{1.0}(Al_{1.0}Si_{2.0})O_6 \cdot 1.00\ H_2O$
1:1:6:2.7	$Na_{0.75}(Al_{0.75}Si_{2.25})O_6 \cdot 1.00\ H_2O$

The parentheses denote equivalent structural positions. The substitution scheme for the high-silica analcites is, therefore, Si→NaAl. It should be noted that the maximum number of sodiums permitted in the structure is 1.50 and the maximum number of H_2O groups is 1.00, although some additional H_2O groups may occupy vacant sodium sites. The formulas written in the form above are given for each analyzed analcite of table 1 calculated on the basis of 14 equivalents (Stevens, 1946).

The specimens on which the compressibility was determined are described in detail below.

Golden, Colorado (U.S.N.M. 84842).—It is presumed that the specimen was collected from Table Mountain, near Golden, Colo. The crystals were chiseled off the walls of a cavity in a basalt flow. The cavity was lined with a 1-2 mm crust of natrolite needles growing perpendicular to the surface. The analcite grew on the natrolite into well-formed crystals not exceeding 2 cm on the largest dimension. The crystals, colorless and transparent, have an index of refraction of 1.485 ± 0.003 and are isotropic to weakly birefringent. The density at 21°C, an average of those of five crystals, was found to be 2.252 before compression. The fragments on which the compressibility measurements were made were of variable size not exceeding 1.5 cm. The powder x-ray diffraction pattern is similar to those of synthetic and some natural analcites believed to be of the cubic type. All the reflections can be indexed on the basis of a cubic cell having $a_0 = 13.712 \pm 0.005$ Å. The very weak reflection corresponding to (200) found in the Flinders, Australia, analcite by Coombs (1955) is present. The unit cell dimension was measured from the reflection (11, 5, 2)–(10, 7, 1)–(10, 5, 5) using the (131) and (132) reflections of quartz ($2\theta = 83.840°$ and $90.832°$, respectively, for $CuK\alpha_1$) as internal standards. The unit cell dimension for this analcite specimen as well as the others studied is in accord with that of Gruner (1928), Taylor (1930), Konta (1953), Coombs (1955), and others. The cell dimension is slightly smaller than that for pure synthetic analcite as predicted by Saha

(1959, p. 304) for other silica-rich analcites. The density calculated from the unit cell dimension of the Golden, Colo., specimen is 2.266, which is somewhat higher than the measured density.

Wasson's Bluff, Nova Scotia (U.S.N.M. 96299).—The crystals were chiseled off a fragment of wall rock of a cavity in basalt. No other zeolites were observed in the vug. The surface of the crystals was slightly iron stained. The index of refraction was 1.487 ± 0.003 and the birefringence was nil to weak. The largest crystal measured 1.7 cm across; however, only those less than 1.0 cm in diameter were used in the compressibility experiment. The powder x-ray diffraction pattern was similar to that of synthetic and some other natural cubic analcites. A weak reflection corresponding to (200) was observed. The high-angle reflections were somewhat broad; the average unit cell dimension is 13.705 ± 0.005 Å. The average density of the sample at 21°C was 2.234 before compression and 2.235 after compression, and the density obtained by calculation from the unit cell measurement is 2.270.

Seisser Alp, Austria (Harvard Museum 51131).—The sections of core were diamond drilled from a single white crystal approximately 6 cm in diameter. The only contaminant is apophyllite, which is concentrated along fractures through the crystal. It is estimated that there is less than 5 percent apophyllite in the core sections. The index of refraction is 1.486 ± 0.003 and the birefringence, nil to weak. The powder x-ray diffraction pattern is in accord with that of synthetic analcite; a reflection corresponding to (200) was not observed. The unit cell dimension is 13.717 ± 0.005 Å, and the average density of the core at 21° C was 2.198 before compression and 2.222 after compression. The density calculated from the unit cell dimension is 2.264.

Fassa Thal, Austria (U.S.N.M. 82457).—The sections of core were drilled from two single crystals, 4 and 4.5 cm in diameter, which grew side by side on a cavity wall. The cores contain apophyllite (<10%) and small amounts of the altered vug wall rock. The powder x-ray diffraction pattern is similar to that of other cubic analcites with the exception of the presence of a weak peak corresponding to (200). The index of refraction of these analcite crystals is 1.489 ± 0.003 and the birefringence is weak. The unit cell dimension is 13.717 ± 0.005 Å and the average measured density of the core sections at 21°C was 2.093 before compression and 2.226 after compression. The density calculated from the unit cell dimension is 2.264.

RESULTS

The results of the individual measurements of decrease in volume are recorded in table 2 and plotted in figure 2. The relative volume change, $-\triangle V/V_0$, is equal to $(V_{2000} - V_p)/V_0$, where V_0 is the volume at atmospheric pressure, V_p is the volume at pressure P, and V_{2000} is the volume at 2000 atm. The experimental values are well represented by the equation:

$$-\triangle V/V_0 = b(P-2000) + c(P-2000)^2.$$

This equation was fitted to the data by the method of least squares and the values of the constants b and c are recorded in table 2. The values of the decrease in volume calculated by means of the equation are recorded for each

TABLE 2

Results of compressibility measurements of analcite

	Golden, Colorado		Wasson's Bluff, Nova Scotia		Seisser Alp, Austria		Fassa Thal, Austria	
P atm.	$-\frac{\Delta V}{V_o}$ Exp.	$-\frac{\Delta V}{V_o}$ Calc.	$-\frac{\Delta V}{V_o}$ Exp.	$-\frac{\Delta V}{V_o}$ Calc.	$-\frac{\Delta V}{V_o}$ Exp.	$-\frac{\Delta V}{V_o}$ Calc.	$-\frac{\Delta V}{V_o}$ Exp.	$-\frac{\Delta V}{V_o}$ Calc.
10,000	0.0345	—	0.0268	0.0262	0.0291	0.0292	0.0292	0.0294
9,000	315	—	220	222	247	246	258	258
8,000	223	0.0218	176	185	204	204	224	221
7,000	162	168	147	149	162	163	185	184
6,000	119	123	118	115	126	126	149	147
5,000	84	84	84	83	91	90	109	110
4,000	54	50	59	54	58	58	70	74
3,000	29	23	30	26	27	28	33	37
2,000	00	00	00	00	00	00	00	00
1,000	-0.0044	—	-0.0034	—	-0.0041	—	-0.0033	—
V (cc)	9.619		8.577		20.817		17.625	
$b \times 10^{-6}$		1.970*		2.485		2.641		3.673
$c \times 10^{-11}$		27.67*		9.86		12.53		0.108

* 2000 to 8000 atm only

pressure and may be compared with the experimental values. Since only the points from 2000 to 10,000 atm were used in determining the equation, the equation is valid only within those limits of pressure.

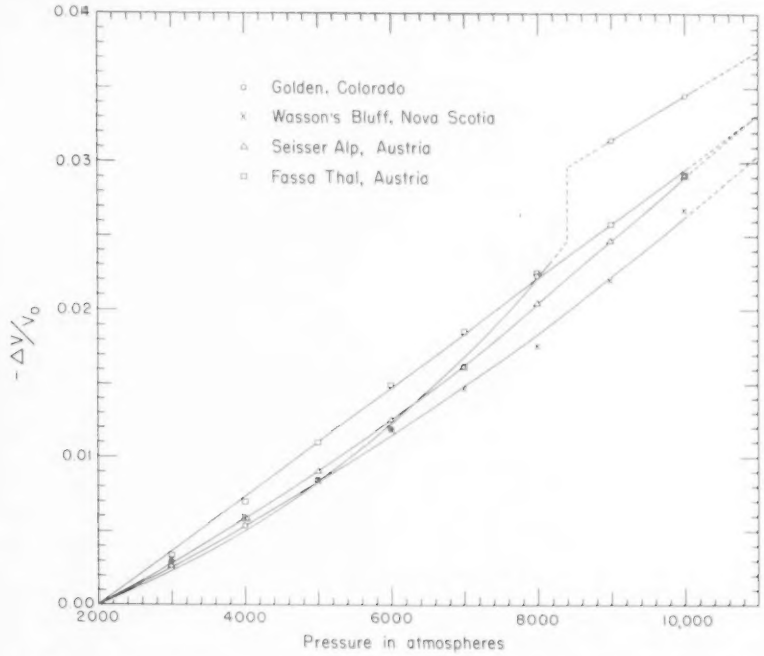

Fig. 2. The relative volume change at 25°C as a function of pressure for four natural analcites. The specimen from Golden, Colo., exhibits a volume change at about 3400 atm. All specimens show an anomalous increase in compressibility with pressure.

The value of the compressibility β which is equal to $-\frac{1}{V_o}\frac{dV}{dP}$ may be obtained by differentiating the above equation:

$$\beta = b - 4000c + 2cP.$$

The specific molal volume change is obtained by integrating the compressibility expression between the limits $P_o = 0$ and $P_1 = P$:

$$\Delta V = -V_o\,[(b - 4000c)P + cP^2].$$

This procedure involves the assumption that the expression for β is valid in the region 0 to 2000 atm. The change in molal volume of the four specimens of analcite (molecular weight = 220.10; densities given above) is given by the following equations for T = 298°K:

Golden, Colo. $\Delta V = -0.843 \times 10^{-4}P - 27.04 \times 10^{-9}P^2$cc/mol

Wasson's Bluff, N. S. $\triangle V = -2.059 \times 10^{-4}P - 9.71 \times 10^{-9}P^2$ cc/mol

Seisser Alp, Austria $\triangle V = -2.120 \times 10^{-4}P - 12.41 \times 10^{-9}P^2$ cc/mol

Fassa Thal, Austria $\triangle V = -3.628 \times 10^{-4}P - 0.107 \times 10^{-9}P^2$ cc/mol

These equations are valid in the region 2000 to 10,000 atm except for the Golden, Colo., specimen, which is valid only in the region 2000 to about 8400 atm. At about 8400 atm the Colorado specimen shows a distinct, reversible volume discontinuity. After compression to 10,000 atm the Colorado analcite crystals were distinctly birefringent and exhibited a large and variable 2V. Macroscopically the fragments were opaque due to a multitude of irregular fractures. There were no observable changes in the index of refraction or x-ray pattern. The other analcite specimens were unchanged after compression.

Several experiments were performed to verify the existence and position of the transition in the Golden, Colo., analcite, and a description of the phenomena involved may be of interest. In routine pressure-volume studies measurements are made on *decompression*, the pressure being reduced in steps of about 1000 atm. At each decompression the contents of the pressure vessel are adiabatically cooled and a time interval of 10-20 min is required at the lower pressure for attainment of thermal equilibrium. During this interval heat flows into the vessel whose contents, because of high piston friction, are maintained at essentially constant volume; as a result the pressure rises. This "pressure regain" depends on factors such as pressure, rate of decompression, and thermal properties, but in general the range rarely exceeds 10-20 atm. Any thermal phenomenon, as for example absorption of latent heat by an inversion, that occurs during decompression produces an anomalous "pressure regain" and is readily observable by means of the pressure gage. The transition in the Colorado analcite was first observed in this manner, with large anomalies being noted on decompressions from 9000-7950 atm.

Some very small discontinuities in volume may be pinpointed by their associated thermal effects. However, if the transition involves a change in volume sufficiently large, an anomalous piston displacement will also be noted in the pressure interval encompassing the inversion. For the Colorado analcite an anomalous piston displacement was observed to be greater between 9000-8000 atm than between 8000-7000 atm. Studies made on decompression in 185 atm steps between 9000-8000 atm (table 3) show that the transition is not as sharp as indicated by figures 2 and 3. It occurs between 8200 atm and 8600 atm with a midpoint at about 8400 atm. Presumably a structurally perfect and chemically pure analcite would invert sharply at about 8400 atm.

It will be noted that the recorded measurements were made on decompression because it is known that metastability is encountered less frequently under these conditions. Observations made on compression, however, showed that the inversion took place readily at about 8400 atm so that the region of indifference is probably small.

Concerning the high-pressure phase of the Colorado analcite it may be said that its compressibility is much lower than any of the low-pressure analcites. The discontinuity in volume, approximately 0.5 cc/mol, cannot be evaluated with great precision since it occurs over a finite pressure range.

TABLE 3

Decompression data for Golden, Colorado, analcite: piston displacements in the pressure region of the polymorphic transition

P, atm (185 atm interval)	Piston Position in inches	Differences in Piston Position in inches
9000	0.1502	
		0.0108
8815	0.1394	
		0.0118
8630	0.1276	
		0.0140
8445	0.1136	0.0133 *
		0.0136
8260	0.1003	
		0.0125
8075	0.0867	
		0.0122
7890	0.0742	
		0.0123
7705	0.0620	
7520	0.0497	

* transition zone.

Volume change and free energy change with pressure for analcite reaction (3)

The volume change for the reaction 2 analcite → nepheline + albite + 2 water at 25°C may be obtained by the following equation:

$$\Delta V = V_{Ne} + V_{Ab} + 2V_W - 2V_{Anal}$$

where the subscript Ne refers to nepheline; Ab, albite; W, water; and Anal, analcite. The volume data for water at 25°C (Adams, 1931, p. 3780), for nepheline and albite (Yoder and Weir, 1951, p. 692), and the present data for the Colorado analcite at 298°K are as follows:

$$V_{Ne} = 54.120 - 1.113 \times 10^{-4}P + 0.28 \times 10^{-9}P^2 \text{ cc/mol}$$
$$V_{Ab} = 100.399 - 2.123 \times 10^{-4}P + 2.17 \times 10^{-9}P^2 \text{ cc/mol}$$
$$V_W = 18.016 - 4.463 \times 10^{-4}P + 14.51 \times 10^{-9}P^2 \text{ cc/mol}$$
$$V_{Anal} = 97.735 - 0.843 \times 10^{-4}P - 27.04 \times 10^{-9}P^2 \text{ cc/mol}$$

The compressibility data for water were fitted by the method of least squares and adequately represent the determined values within the experimental error. The ΔV for the reaction at 25°C is, then:

$$\Delta V = -4.919 - 10.476 \times 10^{-4}P + 85.55 \times 10^{-9}P^2 \text{ cc/mol.}$$

At 2000 atm and 25°C ΔV is equal to −6.672 cc/mol and at 8000 atm and 25°C, −7.825 cc/mol. The important contribution water makes to the volume change is seen in figure 3. The free energy change with pressure at constant temperature is given by the equation:

$$\left.\frac{\partial(\Delta F)}{\partial P}\right|_T = -0.119 - 0.254 \times 10^{-4}P + 2.071 \times 10^{-9}P^2$$

with ΔF in cal/mol, P in atm, and T = 298°K.[3]

[3] 1 cc − atm = 0.024206 cal.

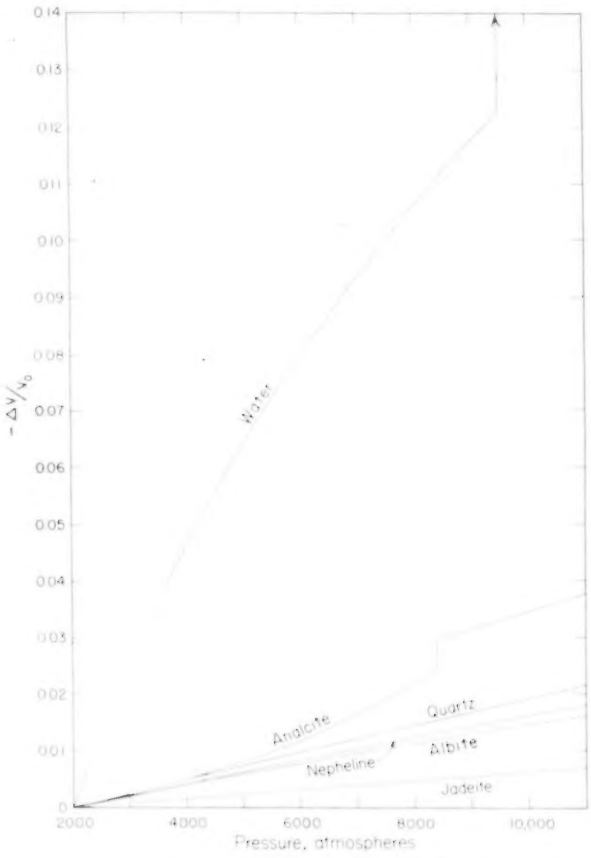

Fig. 3. Relative volume change at 25°C as a function of pressure for phases in the reactions:

2 analcite = nepheline + albite + 2 water
analcite = jadeite + water
analcite + quartz = albite + water

Analcite data from Golden, Colo., specimen; nepheline, albite, and jadeite (Yoder and Weir, 1951); quartz (Adams and Gibson, 1929); water (Adams, 1931).

The analcite reaction takes place at 570°C at 2000 atm water pressure (Yoder, 1950, p. 323). The density of water at this pressure and temperature is 0.6294 (Kennedy, 1950, p. 561) and the molal volume is 28.624 cc/mol. If the thermal expansions of analcite, nepheline, and albite are assumed to be essentially the same, then the reaction would take place with a positive volume change under these conditions. This is the normal volume change in dehydration reactions.

Volume change and free energy change with pressure for analcite reaction (4)

The volume change for the reaction analcite = jadeite + water at 25°C may be obtained from the following equation:

$$\Delta V = V_{Jd} + V_W - V_{Anal}$$

where the subscript Jd refers to jadeite. The volume data for jadeite are given by Yoder and Weir (1951, p. 691) and the others are as indicated above. The ΔV for the reaction at 25°C is, therefore:

$$\Delta V = -19.289 - 4.069 \times 10^{-4}P + 41.42 \times 10^{-9}P^2 \text{ cc/mol.}$$

At 2000 atm and 25°C ΔV is equal to −19.937 cc/mol, and at 8000 atm and 25°C, −19.893 cc/mol. The change for free energy with pressure for the reaction is:

$$\left.\frac{\partial(\Delta F)}{\partial P}\right|_T = -4.669 - 0.098 \times 10^{-4}P + 1.003 \times 10^{-9}P^2$$

with ΔF in cal/mol, P in atm, and T = 298°K.

The reaction was studied experimentally by Griggs and Kennedy (1956) (see also Robertson, Birch, and MacDonald, 1957) and dT/dP was found to be negative. Since the volume change and the entropy change are both negative, Fyfe and Valpy (1959) suggested that the slope of the pressure-temperature curve should be initially positive. The present data, taking into account possible pressure transitions, support in part this suggestion.

Volume change and free energy change with pressure for analcite reaction (5)

The volume change for the reaction analcite + quartz = albite + water at 25°C may be obtained from the following equation:

$$\Delta V = V_{Ab} + V_W - (V_{Anal} + V_{Qtz})$$

where the subscripts Ab and Qtz refer to albite and quartz respectively. The volume change for quartz is given by Adams and Gibson (1929, p. 717, table 3), and the others are as indicated above. The ΔV for the reaction at 25°C is, then:

$$\Delta V = -1.950 - 5.180 \times 10^{-4}P + 43.24 \times 10^{-9}P^2 \text{ cc/mol.}$$

At 2000 atm and 25°C the ΔV is −2.813 cc/mol, and at 8000 atm and 25°C the ΔV was found to be −3.327 cc/mol. The change of free energy with pressure for the reaction is:

$$\left.\frac{\partial(\Delta F)}{\partial P}\right|_T = -0.047 - 0.125 \times 10^{-4}P + 1.047 \times 10^{-9}P^2$$

with ΔF in cal/mol, P in atm, and T = 298°K.

The reaction was investigated by Ellis (1958) and by Valpy (1958) and equilibrium was found, for example, at 285°C and 1000 atm (Fyfe and Valpy, 1959, p. 317; Fyfe, Turner, and Verhoogen, 1958, p. 168, 176). The entropy change of the reaction at 298°K is approximately −7.4 cal/mol/degree using $S_{Ab} = 49.2$, $S_{Qtz} = 10.0$ (Kracek et al., 1951), $S_{Ice} = 9.4$ (Latimer, 1951),

and $S_{Anal} = 56.0$ (King, 1955). The initial slope of the pressure-temperature curve is, therefore, expected to be positive. This conclusion is in accord with the experimental observations.

High-pressure polymorph of analcite

In all four samples of analcite the compressibility is anomalous, that is, the compressibility increases with increasing pressure. This is clearly recognized in figure 2 and by the negative sign of the P^2 term of the molal volume change equations. Such behavior is known in relatively few substances; SiO_2 glass, cerium, and chromium (Bridgman, 1931, p. 168; 1946, p. 22). In cerium the anomalous compressibility increase terminates at a transition to a more stable, high-pressure phase. A transition was realized in the Golden, Colo., analcite, and it is believed that the remaining three samples would invert to a denser form at pressures somewhat greater than 10,000 atm. The failure of the latter samples to invert at the same pressure may be attributed to impurities, as was demonstrated in the case of impure cerium (Bridgman, 1931, p. 236). The representative chemical analyses show significant differences in the composition; the specimens from Golden, Colo., appear to approach most closely the ideal composition.

Of particular interest is the probable structure of the new, high-pressure form of analcite observed in the Golden, Colo., specimen. The transition between the low- and high-pressure forms is reversible and apparently nonquenchable; therefore, direct observation of the structure of the high form is possible only above 8400 atm.

An x-ray powder pattern of the high-pressure form of analcite at 10,000 bars was made using a beryllium bomb similar to that designed by Lawson and Riley (1949, p. 15, fig. 4). The patterns obtained, however, were not of sufficient quality to permit observation of detailed changes, if any, from the low-pressure form. The shattering of the crystals observed after compression constitutes additional evidence that the volumes of the two forms are quite different. The increase in birefringence of the crystals after compression also suggests either that the crystals retained some of the strain induced by the compression or that a possible lower symmetry of the high form was retained. The unit cell size and the character of the x-ray powder pattern obtained at atmospheric pressure on the sample after compression were unchanged. Careful attention was given to the (332), (422), (440), and (640) peaks, which have been observed to split in the birefringent Låven, Norway, analcite (Coombs, 1955, p. 703). Since the increase in birefringence was not observed in those specimens which did not invert, the effect must be associated with the inversion. It is not uncommon for the higher pressure modification of a substance to have a lower symmetry (Bridgman, 1931, p. 248-249).

Many workers have suggested that the birefringence observed in natural analcite crystals can be attributed to internal lattice strain connected in some way with the water content or to twinning of individuals of lower symmetry (Taylor, 1930, p. 12-13). On the other hand, some natural birefringent analcites contain the requisite amount of water and are not twinned (Stewart, 1941, p. 7). One natural analcite from Brewster Co., Texas (Lonsdale, 1928;

Milton, 1936), was believed by A. N. Winchell to have an x-ray powder "pattern which does not correspond to an isometric structure but is not greatly different from one." On the prospect that the Texas analcite might have the high-pressure form, having failed to invert for some unknown reason, material from the same locality collected by Dr. J. T. Lonsdale was examined. The powder x-ray diffraction pattern was similar to that of the Flinders type with spacings consistent with cubic symmetry and having a very weak reflection corresponding to 200. Until the structure of the high form is established the birefringence of the compressed analcite can be attributed to either the strain induced by the volume discontinuity or the retention of possible lower symmetry of the high form not readily observable in x-ray powder patterns. Agents other than pressure may also account for the birefringence in some natural analcites.

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